TECTONICS OF THE NEW GUINEA AREA

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INTRODUCTION

New Guinea is a very large island (2500 km long by 600 km wide) whose
southern half geologically forms the northern part of the Australian con-
tinent. The long fold-and-thrust mountain range, rising to heights above
4000 m, that extends along the spine of central New Guinea is the result
of collisions of this northern margin of Australia with island arcs during
the northward drift of Australia by some 3000 km during the last 37 m.y.
The mountainous and hilly part of northern New Guinea exposes igneous
and metamorphic rocks (Milsom 1985), representing accreted parts of
these collided arcs, which are overlain by younger episutural, or successor,
basins (Figure 1).

The complex system of island arcs that sweep around the eastern side
of Australia and continue around northern Australia–New Guinea, linking
the southwestern Pacific region with eastern Indonesia and the Indian
Ocean, essentially result from interactions between the WNW-moving
Pacific plate and the NNE-moving India-Australia plate (Figure 2). The
Cenozoic fold-and-thrust mountain belts, which characterize the Aus-
tralian continental margin in New Caledonia, New Guinea, and eastern
Indonesia, record the collisions (Figure 1) during the last 40 m.y. (most
occurred during the last 25 m.y., some in the last 3 m.y.) associated with
the continental margin and island arcs marginal to this continent over a
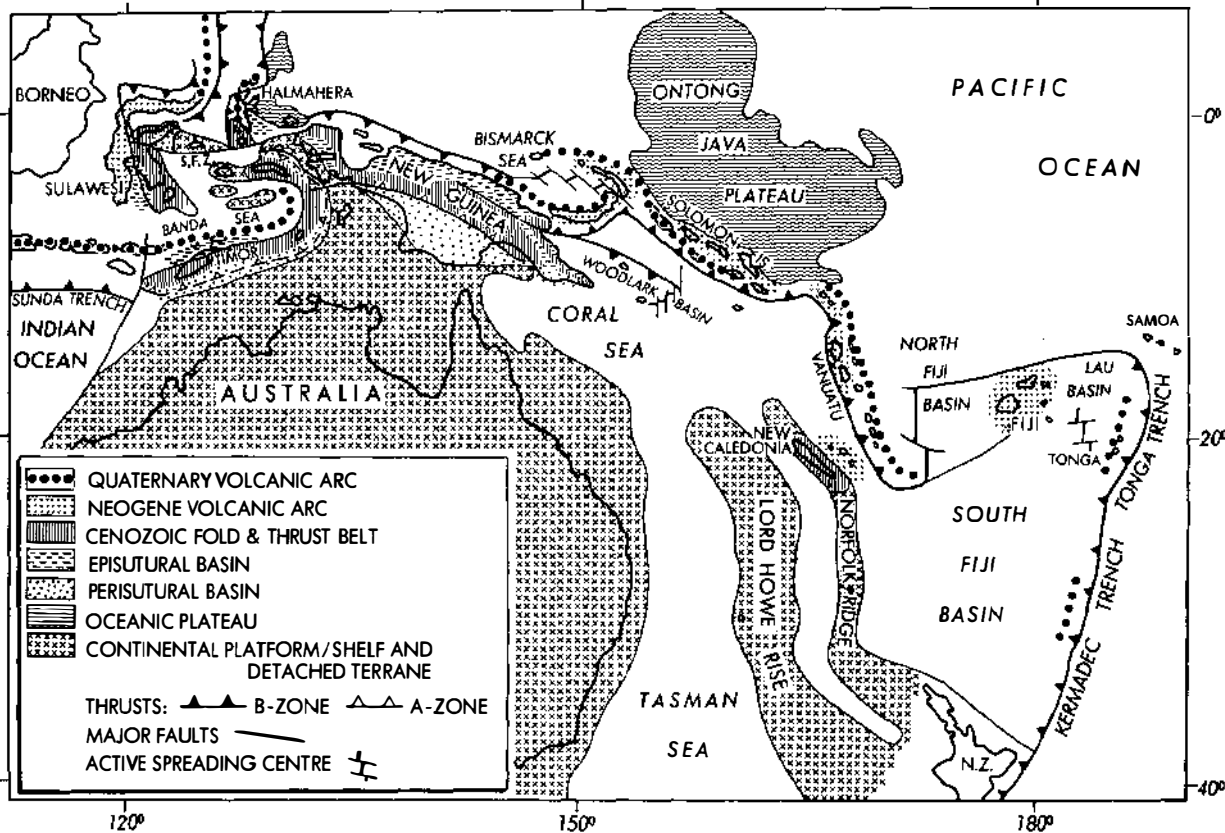strike length of 15,000 km.

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Figure 1   Island-arc systems linked to New Guinea and associated basins. Abbreviation: S.F.Z., Sorong fault zone.
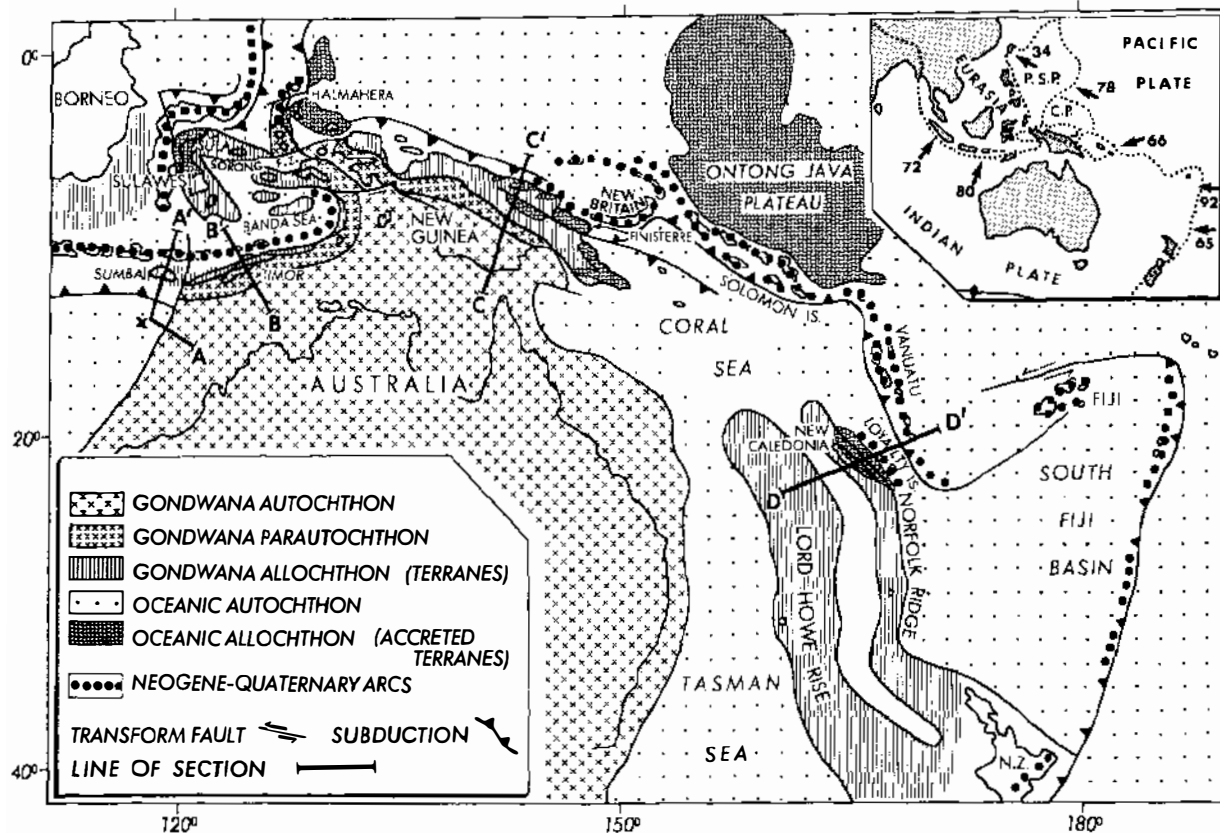
Figure 2 Geological provinces of the New Guinea area. Inset shows present plate configuration and relative motions (in millimeters per year) from Chase (1978). Abbreviations: P.S.P., Philippine Sea plate; C.P., Caroline plate.

The crustal thickening associated with the formation of these fold-and-thrust mountain belts (Chamalaun et al 1976) has led to perisutural basins forming between the fold belts and the Australian craton in eastern Indonesia and southern New Guinea. These peripheral foreland basins (Dickinson 1974) are related to the A-subduction zones of Bally & Snelson (1980).

An account of the tectonics of the New Guinea area must include the tectonically linked active regions of the southwest Pacific and eastern Indonesia (Figure 1). Tectonics is a global phenomenon, so that to some extent the boundaries of New Guinea tectonics are arbitrary. For the purposes of this review, the limits are drawn between the parallels of 5°N and the Tropic of Capricorn and between the meridians of 118°E and 170°W.

Previous Reviews

Hilde et al (1977) interpreted the plate tectonic evolution of the whole area under discussion, together with much of the central Pacific, mainly from the point of view of plate boundary evolution and plate kinematics. Wells (1989) similarly attempted to elucidate the Ontong Java plateau–Solomon Islands evolution. Hamilton (1979) provided a major review covering most of the area under consideration, omitting only the southeast part of New Guinea and the southwest Pacific region. His work remains the most thorough presentation of available data, with abundant maps and an extensive bibliography. The tectonics of New Guinea and the southwest Pacific was reviewed by Kroenke (1984), who omitted consideration of the links with eastern Indonesia. Daly et al (1987) discussed the Cenozoic plate tectonics of New Guinea, the Philippines, and Indonesia with reference to the movements of the India-Australia plate and the mainland southeast Asia portion of the Eurasia plate, but they did not include the important links with the southwest Pacific and the eastern margin of Australia. Bowin et al (1980) analyzed geological and geophysical data for the Banda arc region. Several reviews have focused on New Guinea or parts of that island. In particular, Dow (1977) provided a well-illustrated geological synthesis of Papua New Guinea; Milsom (1985) dealt with the geology of all of New Guinea and the western Melanesian arcs; and Pigram & Davies (1987) considered the island of New Guinea from a mainly stratigraphical viewpoint, reconstructing its paleogeographical evolution with particular reference to the concept of the accretion of tectonostratigraphic terranes. The latter authors emphasized the post-Jurassic evolution of New Guinea from a rifted continental margin to a zone of complex terrane accretion under the influence of oblique plate convergence.

Tectonic Maps

The American Association of Petroleum Geologists (1981) plate tectonic map of the southwest quadrant of the Circum-Pacific covers the whole region under discussion. Hamilton's (1978) tectonic map of Indonesia omits southeastern New Guinea and the southwest Pacific and so excludes those important links with New Guinea. Dupont & Recy (1980) published a tectonic map of the southwest Pacific, but nearly all of New Guinea was omitted from their map. New Guinea is divided politically along a north-south line near the middle of the island, with Irian Jaya occupying the western half and Papua New Guinea the eastern part. Geological maps of the two parts of the island published on a scale of 1 : 1,000,000 (Bureau of Mineral Resources 1972, Dow et al 1986) provide a regional geological basis from which to consider the tectonics.

Geographical Features

The main geographical features (Figure 1) in the region under consideration, which has about the same area as the continental USA, are (a) the wide continental shelf linking northern Australia with the island of New Guinea and extending westward into the Timor Sea and eastward into the Coral Sea plateau; (b) the mountainous island of New Guinea; (c) the Banda arc of eastern Indonesia, with its deep marine troughs locally descending to 7000 m, and the associated Banda Sea back-arc basin; (d) the island arcs of the southwest Pacific (Bismarck, Solomon, Vanuatu, New Caledonia, Fiji, and Tonga being the principal groups), with trenches locally descending to more than 9000 m and with associated back-arc basins; (e) the long, narrow, mainly submarine ridges of thin continental crust (Lord Howe Rise and Norfolk Island Ridge) off eastern Australia and their associated deeper marine troughs of the Tasman Sea.

Geophysical Features

The outstanding geophysical feature of this area is the very high level of seismicity in the numerous Wadati-Benioff zones (Figure 1) that characterize both southwest Pacific and eastern Indonesian tectonics (Hamilton 1978, American Association of Petroleum Geologists 1981). There is also a detectable seismic expression of many important strike-slip faults associated with the active deformation across the region. The strong link between the Wadati-Benioff zones and arc volcanism was discussed by Hatherton & Dickinson (1969).

This region has provided seismic and other geophysical data for some of the most important analyses used for interpreting major global tectonic phenomena. For example, the paper by Sykes (1966) on the seismicity of

the Tonga region demonstrated the geometrical relationships of the narrow, steeply dipping zone of earthquakes to surface expressions of trench and arc. Oliver & Isacks (1967) used seismic data from the Tonga Trench region, especially aspects of body-wave amplitudes from deep events, to interpret the Wadati-Benioff zone as one of high Q. From this they proposed that the inclined earthquake zone represents a dipping, cool slab of Pacific lithosphere. Earlier, Vening Meinesz (1948), who had measured gravity from a submarine off eastern Indonesia, calculated the isostatic anomalies of the region, which showed how the trenches and the perisutural, proximal foreland basins are characterized by strongly negative isostatic gravity anomalies.

Field Laboratory for Active Tectonic Processes

The New Guinea area, embracing eastern Indonesia and the southwest Pacific, is the world's outstanding field laboratory for studying active tectonic processes. The tectonic consequences of convergent plate movements are exceptionally well displayed here. Volcanic arcs expose a very wide range of development. Active spreading in back-arc basins is known from the Lau, North Fiji, and other basins. Episutural basins are actively developing on the orogenic hinterlands. Perisutural foreland basins (Figure 1) are actively growing at the cratonic margins of the collisional orogen. In no other part of the world is the transition from rifted to active continental margin so well displayed. The concept of tectonostratigraphic terranes can be investigated where these terranes are actively forming and accreting. Here we can study tectonic processes before erosion removes the highly fossiliferous and hence datable sediment cover from which rates of processes may be determined in the orogenic hinterland, and before overprinting by later tectonic or thermal events obscures the original features, a problem that bedevils the analysis of many older orogens.

STRATIGRAPHY AND STRUCTURE

Main Geological Provinces

At one level of discussion it might be said that only two geological provinces are involved: (a) Gondwana and its various fragments, some of which have been detached and later accreted onto the large Gondwana fragment of Australia–New Guinea; and (b) the oceanic province with its younger volcanic crust, ophiolite fragments, and island-arc magmatic products, some of which have been accreted to parts of Gondwana and some of which still remain part of the oceanic realm. Figure 2 attempts to identify the distribution of these two provinces. Discussion of geological provinces across this large, complex area at a more detailed level is difficult and

controversial. It revolves necessarily around stratigraphical, faunal, and floral affinities of cover-rock sequences and involves perceptions of petrological and metamorphic provinces; in addition, it relies in part upon paleomagnetic measurements and radiometric dating from blocks of continental, oceanic, and island-arc rocks.

STRATIGRAPHY OF GONDWANA PROVINCE   The stratigraphical successions of the parts of this area that can be ascribed to a Gondwana continental origin are of two kinds: (*a*) autochthonous, or parautochthonous in the case of the strongly deformed zones, especially the fold-and-thrust belts; (*b*) allochthonous, in the sense that such sections no longer enjoy their original relationships with their basement, having been mechanically detached in the deformation process. The allochthonous sections belong for the most part to those fragments of Gondwana that were detached from Australia–New Guinea during Late Jurassic and Eocene rifting of northeastern Australia–New Guinea and from the eastern margin of Australia during Paleocene rifting and spreading. Some authors have applied the term tectonostratigraphic terrane to the continental crustal (or lithospheric) blocks removed from Gondwana by rift-drift processes and accreted later to this margin. The stratigraphy of these two types of Gondwana fragments, distinguished on the sketch map (Figure 2), is summarized in Figures 3 and 4. Within the area under consideration, geological subprovinces have usually been defined on the basis of present-day configurations—for example, (*a*) the Banda arc, (*b*) Sulawesi, (*c*) the Moluccas, (*d*) New Guinea, and (*e*) New Caledonia. This has been the basis for stratigraphical and structural considerations (van Bemmelen 1949, Hamilton 1979, and many others). The binary classification of geological provinces employed here is based on stratigraphy.

The Gondwana cratonic basins dominated the stratigraphy of New Guinea, the Moluccas, and the Banda arc until the major Jurassic rift-drift phase produced the ingression of Mesozoic Tethys along the newly rifted continental margin of northeast Australia–New Guinea (Figure 3). Cretaceous–Paleogene history is mainly one of a subsiding rifted continental margin influenced by global sea-level changes and local spreading events until the arc-continent collisions begin to affect New Caledonia in the Eocene, central New Guinea from the Oligocene onward (Pigram & Davies 1987), and the northeast Australian margin from the Miocene onward, giving rise to the Banda arc (Audley-Charles 1987) and Sulawesi (Simandjuntak 1986) orogenic belts. In all of these areas of arc-continent collision, nappe piles built up and moved toward the Australian craton (Audley-Charles 1986), engendering the downwarping of the crust and development of perisutural basins or proximal foredeeps (Jordan 1981).
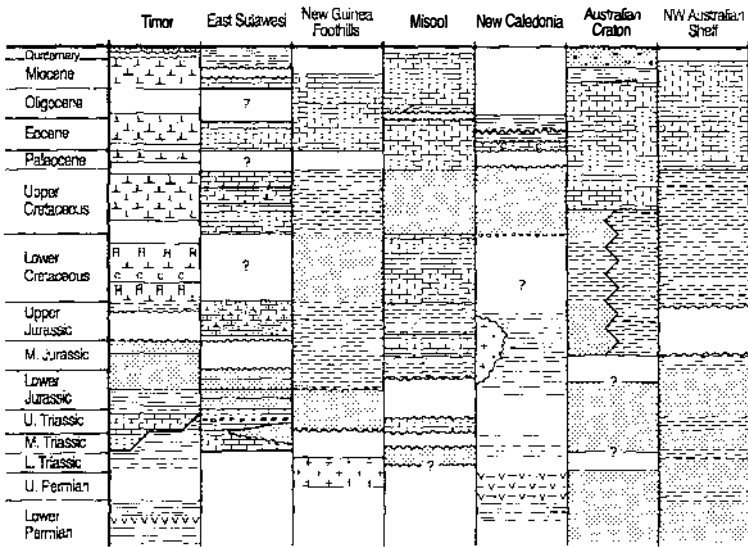
Figure 3   Stratigraphic summary of the autochthonous-parautochthonous Gondwana fragments. See Figure 4 for explanation of rock symbols.
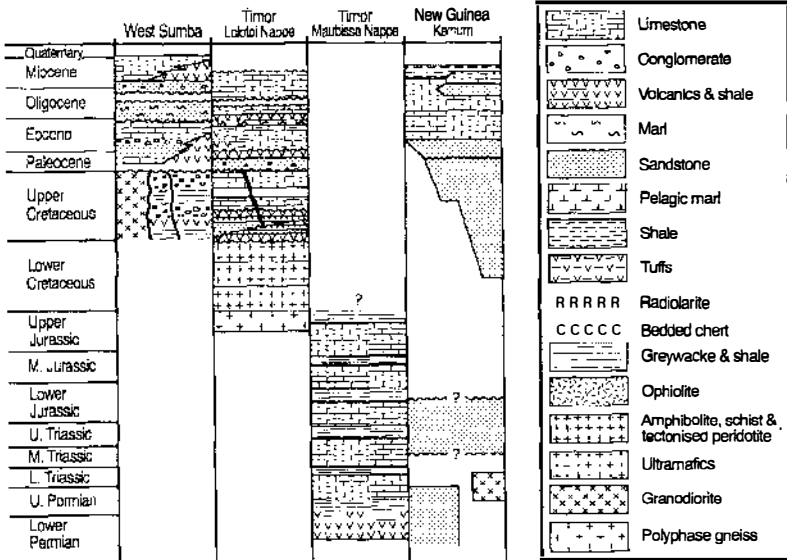

Figure 4   Stratigraphic summary of the allochthonous Gondwana fragments (terranes).

Figure 4 reveals the continental affinity of the allochthonous elements that developed in an intra-Tethys or intra-Pacific oceanic environment after their rifting from Australia–New Guinea.

STRATIGRAPHY OF THE OCEANIC PROVINCE   The stratigraphical successions whose origin can be ascribed to the oceanic realm are of two types: (*a*) those that remain as part of the oceanic realm; and (*b*) fragments of seamounts, oceanic plateaus, and ocean-floored volcanic arcs accreted to parts of Gondwana, which are regarded by Pigram & Davies (1987) as tectonostratigraphic terranes. The stratigraphy of these accreted fragments (terranes) is summarized in Figure 5. The widespread development of Paleogene arcs and back-arcs perhaps resulted from the large-scale plate reorganization and movements (Gordon & Jurdy 1986, Daly et al 1987, Harris 1989). Continuing plate convergence, particularly between the three large plates of southeast Asia, India-Australia, and the Pacific during the Neogene–Quaternary has led to active arc configuration.

PLATE TECTONIC CONFIGURATION AND GEOLOGICAL STRUCTURE   The American Association of Petroleum Geologists (1981) map of the southwest quadrant of the Circum-Pacific provides a good summary of the tectonic
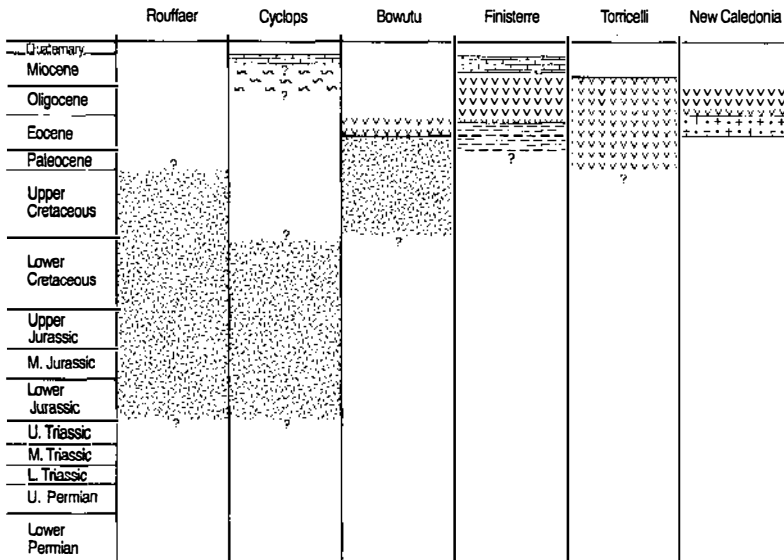
Figure 5   Stratigraphic summary (after Pigram & Davies 1987) of the accreted oceanic fragments (terranes) in northern New Guinea and in New Caledonia. See Figure 4 for explanation of rock symbols.

plate configuration and vectors for the whole area. There are five major plates involved (Figure 2), and a number of smaller plates have been identified, especially at the boundary between the India-Australia plate with the Pacific, Caroline, and Philippine Sea plates. Detailed local studies (Hall & Nichols 1990) have suggested that the boundaries of some of these minor plates are highly unstable. The geological structure of the area is summarized by a series of sections drawn across the different sectors (Figure 6) along the whole 15,000-km length of the deformation zone at the northeast and eastern margin of Australia-New Guinea. The regional geological context of these cross sections is illustrated in Figure 2.

OUTLINE OF GEOLOGICAL HISTORY

Precambrian

The geological core of the area is the Precambrian shield of Australia, which may continue northward (unexposed) as part of the crystalline continental basement below parts of southern New Guinea. The presence of Precambrian rocks in the basement of New Guinea and in the islands with continental basement in eastern Indonesia and the southwest Pacific, which are linked tectonically to the New Guinea area, has neither been established nor disproved (Hamilton 1979, Veevers 1984, Kroenke 1984).

Paleozoic

The occurrence of lower Paleozoic rocks in New Guinea (Milsom 1985) and the other islands under discussion appears to be rare. Silurian graptolites have been found in place in western Irian Jaya (Pieters et al 1983), and Silurian fossils were reported from limestone pebbles in the float of streams draining the southern slopes of the highlands of eastern Irian Jaya in the central orogenic fold belt (Visser & Hermes 1962). Upper Paleozoic rocks are widely exposed in the Irian Jaya part of New Guinea, especially along the southern margins of the central ranges, where they are mainly Permian terrestrial deposits (Visser & Hermes 1962, Dow et al 1986). Fossiliferous marine Permian rocks have been found in the Bird's Head region (the far northwestern part) of Irian Jaya (Archbold et al 1982) and in Timor (Audley-Charles 1988) but not in the other islands of the Banda arc. Similar marine Permian facies are also found in New Caledonia (Paris & Lille 1977). Metamorphic rocks intruded by Permo-Triassic granitoids have been reported from the Sula Islands (Pigram et al 1985). Metamorphic rocks are widely exposed in New Guinea and in the islands of the Banda arc and the southern Moluccas (Barber & Audley-Charles 1976, Pigram & Panggabean 1984), where some protoliths could be early or late Paleozoic in age.
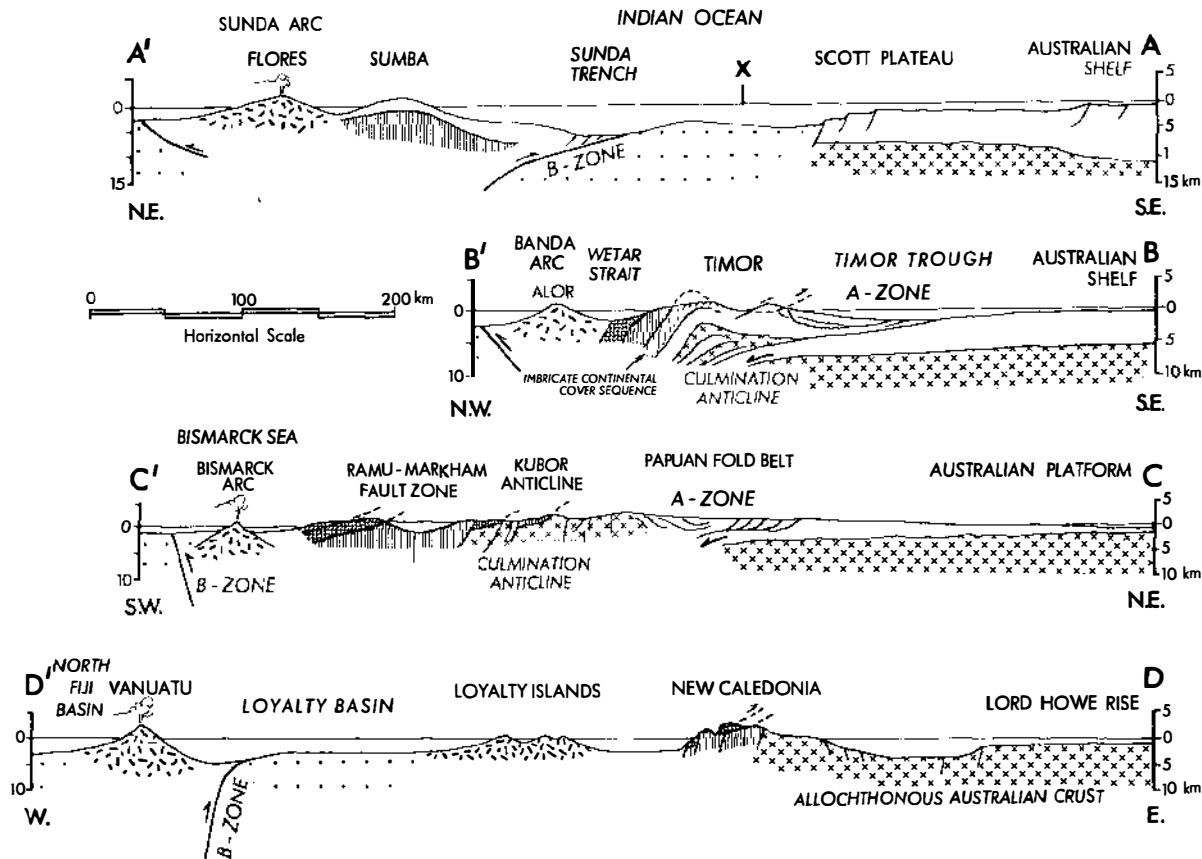
Figure 6 Geological sections across the orogenic belt. See Figure 2 for section locations and explanation of symbols.

Permo-Triassic Paleogeography

The paleogeographical configuration of eastern Gondwana in the Late Carboniferous and Early Permian interprets New Guinea and eastern Indonesia (Metcalfe 1990) as parts of Gondwana cratonic basins (Figure 7). The islands of the southwest Pacific with continental basement, such as New Caledonia and North Island of New Zealand, seem most likely to have formed part of the active continental margin of Australian Gondwana in the late Paleozoic (Kroenke 1984). The Triassic and Early Jurassic are characterized by Gondwana cratonic basin facies in much of New Guinea and eastern Indonesia (Audley-Charles 1988). This is a fully marine facies in eastern Indonesia but gives way to dominantly terrestrial facies of Permo-Triassic age along the southern margins of the central ranges in New Guinea (Milsom 1985). Marine Permo-Triassic and Early Jurassic
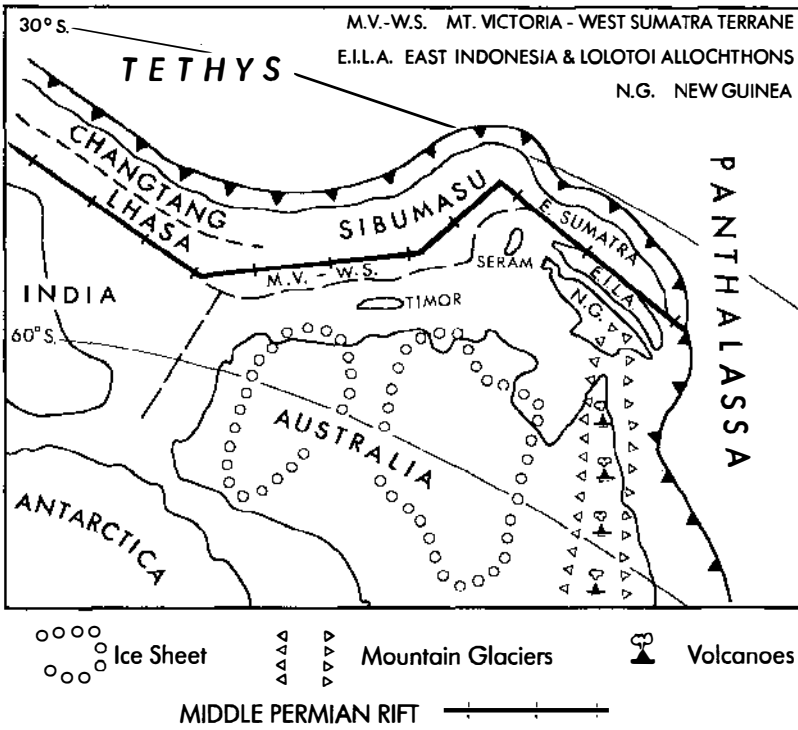
Figure 7 Early Permian paleogeography, from Metcalfe (1990) and Audley-Charles (1987).

deposits occur in the Bird's Head region of Irian Jaya, where they may be correlated with the parautochthonous successions of the Banda arc (Figure 3).

Mesozoic Rifted Continental Margins

The continental margin of northern Australia–New Guinea in the Mesozoic (Figure 8) has been defined with broad agreement (Hamilton 1979, Veevers 1988). Where different interpretations exist (Pigram & Panggabean 1984), they stem largely from the problem of identifying the boundary of the Australian continental margin in the complex fold-and-thrust belt. There is general agreement, based upon stratigraphical and structural observations made in the field and on oil company multichannel seismic reflection profiles, that this continental margin was created as an Atlantic-type rifted margin in the Jurassic.

The eastern margin of Australia during the Mesozoic was an active one, along which andesitic volcaniclastic sediments accumulated (as, for example, in New Caledonia). The subduction zone is thought to have

E.I.A. EAST INDONESIA
        ALLOCHTHONS

L.A. LOLOTOI
        ALLOCHTHON

S. SERAM

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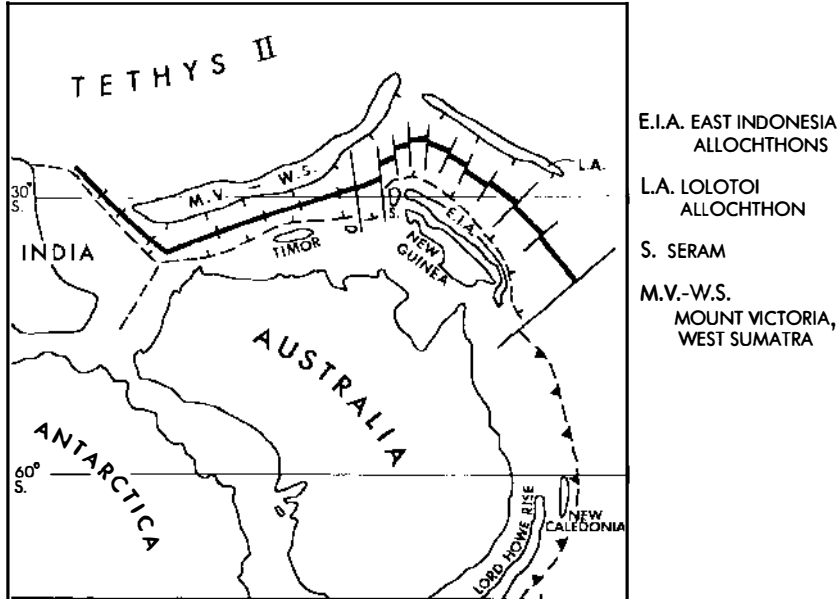   MOUNT VICTORIA,
        WEST SUMATRA

Figure 8   Late Jurassic paleogeography, from Meltcalfe (1990), Audley-Charles (1987), and Harris (1989).

dipped toward the Australian continent (Blake et al 1977) until the Early
Cretaceous, when volcanism ceased. An ophiolite was emplaced on the
axial part of New Caledonia in the Cretaceous (Paris & Lille 1977). The
Late Jurassic–Early Cretaceous metamorphic event is correlated with the
uplift of New Caledonia prior to its separation from the Australian con-
tinent by spreading of the Tasman Sea in the Paleocene.

There has been some controversy concerning the identification of the
continental blocks removed from the Australian part of eastern Gondwana
during the Jurassic. Audley-Charles (1988) suggested that the detached
continental blocks were parts of Burma, western Thailand, Malaya, and
Sumatra. Metcalfe (1990) has argued cogently that, on the contrary, those
blocks were removed from eastern Gondwana in a mid-Permian rifting
episode, and he has gone on to show that the continental blocks removed
from central New Guinea and northeast Australia to create the Jurassic
rifted continental margin there are probably parts of western Burma and
western Sumatra, together with what are now various allochthonous
elements found in the Banda arc collision zone. Metcalfe's interpretation
of the rifted margin of northeast Australia–New Guinea can be linked
(Figure 8) with the reconstruction of the continental margin of eastern
Australia in the Jurassic (Falvey & Mutter 1981).

Cenozoic Evolution

The main features of the paleogeographical evolution of the New Guinea
area during the Cenozoic follow from major plate developments over the
last 130 m.y. (summarized by Veevers 1984): (a) India was rifted from
Antarctica and western Australia at about 128 Ma, and this event was
followed by spreading that moved India northward, creating the Indian
Ocean. (b) The rifting of Australia from Antarctica during the Cretaceous
(Cande & Mutter 1982) was followed by a long period lasting almost 60
m.y. in which only very slow spreading occurred between Australia–New
Guinea and Antarctica. (c) At 82 Ma, slow spreading began between
eastern Australia and the Lord Howe Rise–New Zealand region, initiating
the Tasman Sea. This was followed by rapid spreading between 66 and 57
Ma, creating the present Tasman Sea. (d) At this time (65–62 Ma) there
was a very short phase of rapid spreading in the Coral Sea (Weissel
& Watts 1979), which separated the Papuan Peninsula from northeast
Australia. (e) At about 43 Ma there was a major change in the motion of
the Pacific plate (Engebretson et al 1985, Gordon & Jurdy 1986) from a
north-northwestward motion to a west-northwest motion. (f) From 37
Ma to the present there has been a phase of rapid spreading between
Antarctica and Australia that has moved Australia about 3000 km to the
north-northeast.

Eocene

During the Eocene, while Australia was being carried northward by spreading on the India-Antarctic Ridge, there may also have been spreading along the northern margin of New Guinea (Kroenke 1974, Gordon & Jurdy 1986), as well as in the region north of what is now northeast Australia (Figure 9). The present Solomon plate (Davies et al 1984), part of the Banda Sea, and the Palelo arc remnant in Sumba-Timor-Seram (Harris 1989) are thought to be Eocene in age and related to this spreading episode at the northeast Australia–New Guinea margin. Pigram & Davies (1987) proposed that the Papuan Ultramafic Belt amalgamated in a com-
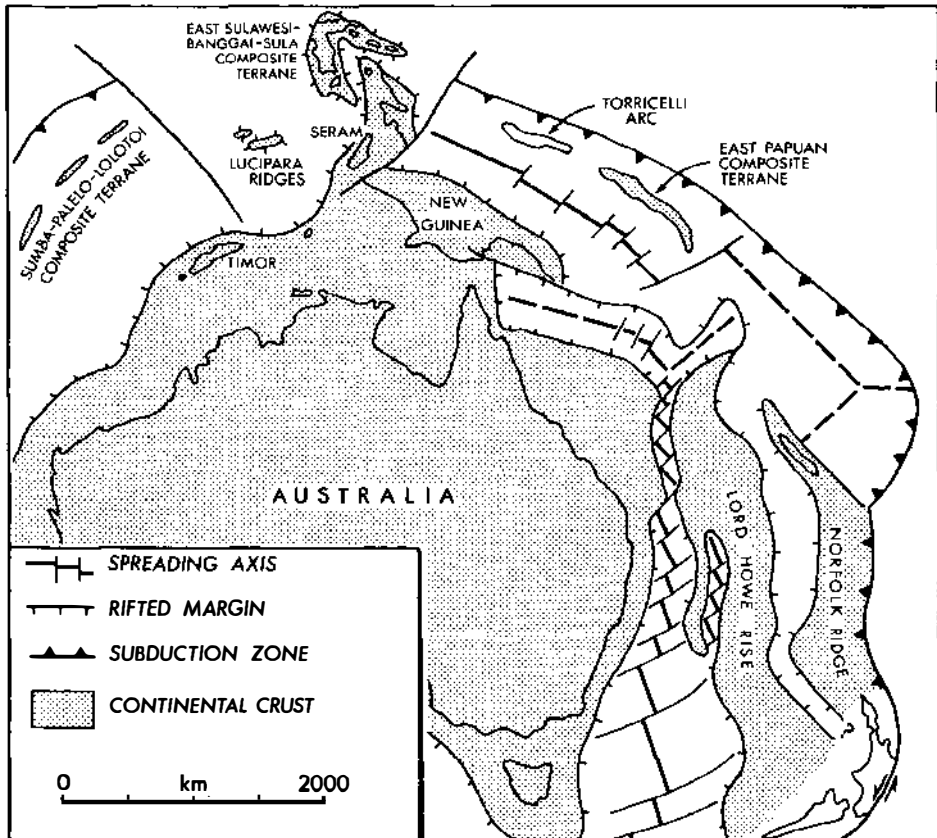
Figure 9 Middle Eocene paleogeography, partly after Falvey & Mutter (1981), Kroenke (1984), and others.

posite East Papuan terrane prior to its Oligocene accretion to the New Guinea margin. Similarly, the Lolotoi-Palelo terrane of Sumba-Timor-Seram in the Banda arc may represent an early Eocene tectonic amalgamation (Harris 1989) prior to the Pliocene emplacement of the Lolotoi-Palelo nappe on the Australian continental margin. Two further examples of amalgamating arc terranes at this time are seen in the Eocene volcaniclastic-carbonate facies of Halmahera in the Moluccas (Hall & Nichols 1990) and in New Caledonia. In the latter case oceanic crust in the late Eocene was thrust from the northeast over the volcanic arc and the rifted continental margin fragment (Kroenke 1984).

Oligocene

The Oligocene saw a major shift to convergent tectonics in this area resulting from the change in plate movements (Gordon & Jurdy 1986). Volcanic arcs with associated carbonates continued to be active around the northeast margin of Australia–New Guinea. Robinson et al (1976) report 4.5 km of Oligocene volcanic rocks exposed in the Adelbert and Finisterre ranges of Papua New Guinea. The Palelo arc (Figure 9) of the (Pliocene) Lolotoi-Palelo nappe in the Banda arc also continued to be active at this time (Harris 1989).

Miocene

The Miocene witnessed the widespread collision of arc terranes at the continental margin of northeast Australia–New Guinea. These events include the accretion of terranes to what is now northern New Guinea (Pigram & Davies 1987), the accretion of east Sulawesi-Banggai-Sula to western Sulawesi (Audley-Charles 1987), and the beginning of the collision of northeast Australia with the Banda arc (Figure 10). Pigram & Davies (1987) consider that all of the Bird's Head region of New Guinea, including Misool, is composed of terranes that accreted to the Australia–New Guinea margin during the Mio-Pliocene. It is possible that the stratigraphical variations between most of the southern part of the Bird's Head and Misool region and the rest of New Guinea could be accounted for as changes along strike of the Australia–New Guinea Gondwana craton and later along its rifted margin.

The Mio-Pliocene converging interaction of the Pacific and India-Australia plates with each other and with the other plates to the north (the Caroline, Philippine Sea, and Sunda plates) created Wadati-Benioff zones and their associated volcanic island arcs (Figure 2). The continental margin of northern Australia–New Guinea began to collide with trenches and associated forearcs. The Miocene was a time of considerable volcanism in the oceanic realm that was associated with new arcs [e.g. Fiji, Loyalty, and
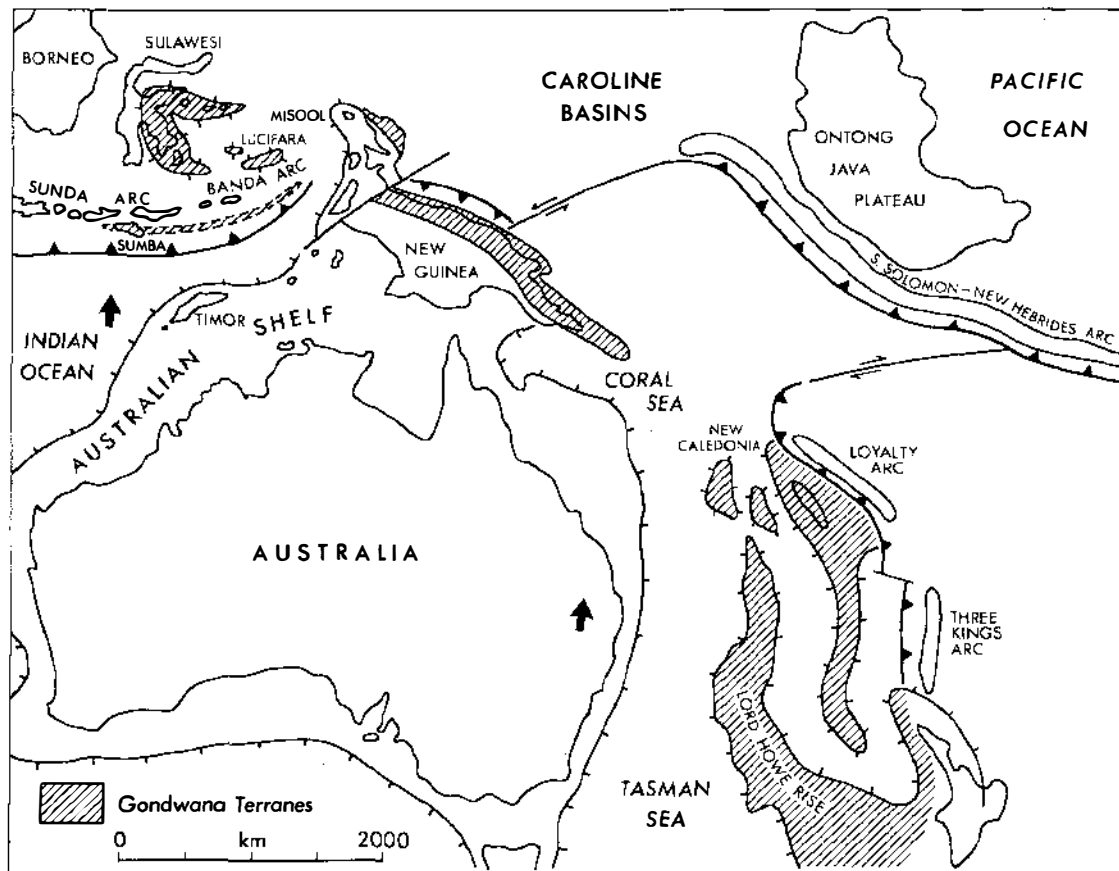
Figure 10 Late Miocene paleogeography, partly after Kroenke (1984) and others.

New Britain (Kroenke 1984)]. Various major extensional and strike-slip features (American Association of Petroleum Geologists 1981) have developed in conjunction with these complex plate movements and related tectonics. The Sorong fault zone is a major strike-slip crustal suture that was active in the late Miocene. A feature of the Miocene was the widespread accumulation of thick shallow marine carbonate sediments associated with volcanic terranes, such as Finisterre and New Britain (Milsom 1985), Halmahera (Nichols et al 1990), and Sumba-Palelo-Lolotoi (Harris 1989).

Pliocene

The Pliocene was a time of major deformation and uplift along the 15,000 km of the orogenic zone marginal to the northeastern and eastern margin of Australia. Episutural basins developed on the orogenic hinterland of New Guinea, on Halmahera, and on the arc-continent collision zone in the Banda arc. The Pliocene also marked the development of the perisutural foreland basins (Figure 1). The clastic fill for these basins had its source in the rising fold-and-thrust mountain chains. Terrane accretion was active during the Pliocene. The Finisterre terrane had accreted to New Guinea by about this time. It passes eastward into the still active New Britain arc.

Within the compressional megasuture (Bally & Snelson 1980), extensional forces were locally active. For example, two oceanic spreading centers opened: the Woodlark system (Milsom 1970, Weissel et al 1982) and the Bismarck system (Taylor 1979) at about 3.5 Ma. Cullen & Pigott's (1989) speculation linking the northern end of the Ramu-Markham fault (Figure 6) throughout northern New Guinea to the Sorong fault zone does not seem justified by the regional mapping (Bureau of Mineral Resources 1972, Dow et al 1986). In the Banda arc region the opening and deepening of the Weber Deep appears to be related to the eastward migration of the arc and forearc by left-lateral strike-slip faulting without seafloor spreading (McCaffrey 1989).

Quaternary

Quaternary geology throughout the area continues the evolutionary trends of the Miocene and Pliocene. In the Banda arc system the tight convergence of Timor with the volcanic arc appears to be related both to the shutting off of volcanic activity there and to the locking of the thrust faults by their rotation into steep angles in the convergent zone, which is now almost aseismic in the Wetar-east Timor sector (Price & Audley-Charles 1987). Strike-slip accommodation motions across the orogenic zone began to play an increasingly important role after collision (Charlton 1989, Harris 1989, McCaffrey 1989).

In the Moluccas, convergence appears to have been influenced by the heterogeneous character of the Philippine Sea plate, so that in the Halmahera region convergence is being transferred from subduction at the Philippine Trench via a NE-SW dextral transpression zone across Halmahera to the Molucca Sea collision zone (Hall & Nichols 1990).

In New Guinea, the main Pliocene tectonic events continue. Hill & Gleadow (1989) have shown that the Papuan fold belt was uplifted, eroded, and cooled from the earliest Pliocene (5 Ma) to the present day. One apparently anomalous development is the calc-alkaline volcanism in the highlands of southeast Papua and the New Britain arc. The highlands are not obviously associated with any existing plate boundary, and this volcanic zone has aroused considerable debate (Hamilton 1979).

The southwest Pacific region continues to be marked by diverse local collisions associated with the complex interacting systems of colliding ridges (e.g. Louisville Ridge–Tonga collision), arcs (Loyalty–New Caledonia), and oceanic plateaus (Ontong Java–Solomon arc). But this part of the area is dominated by the active Pacific subduction zone from Kermadec through the Tonga, Vanuatu, and Solomon arcs (some 7500 km long). It displays a great change of strike between Samoa and Fiji that is associated with the active spreading (Kroenke 1984) in the North Fiji Basin (Figure 1).

ALLOCHTHONOUS TERRANE CONCEPT APPLIED TO THIS AREA

Nur & Ben-Avraham (1982) characterized as allochthonous terranes the many continental fragments derived from Gondwana that are found among the arcs marginal to northern and eastern Australia. Ben-Avraham et al (1981) identified the oceanic plateaus of the region as potentially allochthonous terranes. Silver & Smith (1983) compared these continental fragments, arcs, and oceanic plateaus directly with the Mesozoic North American Cordillera to illustrate the active terrane processes of the New Guinea area. Pigram & Davies (1987) have identified at least 32 terranes in New Guinea, many of them composite. Cullen & Pigott (1989) are more restrained in their designation of terranes, recognizing that some lateral variations in facies and structure may occur within a single terrane or block. However, Audley-Charles & Harris (1990) have shown, on the basis of metamorphic and volcanic history and detailed stratigraphy of cover-rock sequences, that relatively small blocks of continental crust, presently only some tens of kilometers long, are the product of tectonic accretion of what are now called terranes. These terranes appear to amalgamate in an oceanic setting before accretion to the continental margin. They may be

emplaced as roof thrusts of the forearc over the continental margin, after which they may become attenuated and broken up as a consequence of uplift and erosion of the mountain range.

ACTIVE TECTONIC PROCESSES

Mechanisms, derived from theoretical and computer modeling and based partly on data from laboratory experiments, proposed to explain the formation of fold-and-thrust mountain belts and their associated terranes, basins, Wadati-Benioff zones, trenches, and arcs need to be tested in areas of active tectonics. The New Guinea area is probably the best field laboratory for such investigations.

Development of the Orogenic Hinterland

Oblique plate convergence has led to the formation of fold-and-thrust mountain belts on an Alpine-Himalayan scale during the last 5 m.y. in New Guinea, the Moluccas, and the Banda arc. The Plio-Pleistocene uplift of these fold-and-thrust belts from below sea level to mountainous elevations has preserved the highly fossiliferous sediment-cover sequences, even in the orogenic hinterlands. These belts provide exceptional opportunities for studying the structural evolution and rates of deformation of the arc-continent collisional process associated with plate convergence. In most older orogenic belts, the hinterland is generally characterized by major strike-slip faults, multiphase deformation, and reworking by metamorphic and thermal events; it is usually deeply eroded, exposing only the crystalline core. The hinterland usually represents the suture zone between the colliding continent and the magmatic arc. In older mountain belts the hinterlands, which should contain the key to the collisional processes and mechanisms, have generally lost their cover-rock sequences from which the original geometry of the collision zone and the rates of deformation and uplift might have been deduced. One of the consequences of collision-related nappe emplacement is that postcollisional uplift leads to nappe attenuation by low-angle faulting, which in Timor and Seram is accompanied by much steep-angled normal faulting (Harris 1989). Similar processes may be active in the highlands of New Guinea.

Thermal Controls of Arc-Continent Collision Processes

In the Banda arc islands of Timor and Seram, where the arc-continent collision occurred as recently as 3 Ma, the cover-rock sequence of the uplifted hinterland has been preserved from overprinting and erosion, and thus is possible to deduce the geometry of the roof thrust that carried part of the Banda forearc over the colliding Australian continental rise and

slope (Figure 6). The fossiliferous cover-rock sequences have allowed the rates of nappe emplacement and of immediate postcollisional uplift (Audley-Charles 1986) to be measured. Harris (1989) has argued that it was the influence of the colliding north Australian continental margin (being 160 m.y. old and hence geologically cold) that led to its relatively steep subduction, so that part of the Banda forearc overrode the converging continental margin forming the allochthon. This scenario is in strong contrast with the arc-continent collision in Taiwan. There the South China Sea was very young (ca. 15 Ma) at the time the Luzon arc collided with the southwest China continental margin. That margin was therefore much hotter than the north Australian margin, so that southwest China rode over the colliding Luzon forearc and arc, in contrast to the Banda collision geometry.

Rotation of Major Faults in Postcollision Convergence

Price & Audley-Charles (1987) argued that major thrust faults in the Banda collision zone had, during the continuing plate convergence after collision, rotated into steeper angles until they locked. The relatively aseismic condition of the east Timor-Wetar collision zone (McCaffrey 1989) suggests that the steep fault zone in the Wetar Strait, which defines the boundary between the Banda volcanic forearc and the Australian margin, has locked. This conclusion is consistent with the results of a recent GLORIA survey in this strait (Masson et al 1990), which revealed little or no recent faulting along the north margin of east Timor.

Rates of Foreland Thrust Belt Processes

Apatite fission track analyses (Hill & Gleadow 1989) have indicated that the Papuan fold belt was uplifted, eroded, and cooled from the earliest Pliocene (5 Ma) to the present day. Thrusting of the large anticlines at the mountain front probably occurred within the last 1 m.y. Similarly, in Timor the strong folding with thrusting and local imbrication of the Australian continental margin parautochthon, accompanied by the emplacement above of allochthonous nappes that moved 50 km across the deformed Australian margin, was completed within 0.8–0.4 m.y. (Audley-Charles 1986).

Episutural Basin Formation

Pigram & Davies (1987) have pointed out that many of the allochthonous terranes making up the tectonic collage of northern New Guinea are covered with sedimentary basins of Miocene and younger age. These basins are typically filled with 3–7 km of turbidites. Some appear to be pull-apart in origin, related to the postcollisional extension seen in many orogens;

others may be related to strike-slip faulting, while some basins (such as the Ramu-Markham basin) are of uncertain origin. The northern Moluccas in the region of Halmahera, although a much younger (late Pliocene–Quaternary) fold-and-thrust belt, is also developing episutural basins, which appear to be related to major strike-slip movements of splays of the Sorong fault zone (Hall & Nichols 1990). Some different kinds of episutural basins, such as the Wetar Strait and Weber Basin, are forming on the collision zone in the Banda arc (Figure 1) above what was the precollision forearc basin. Where the Australian continental margin has ridden over part of the forearc in the Timor region, the Wetar Strait appears to be evolving as a basin on what remains of the forearc, here reduced to 50 km or less in width.

Perisutural Foreland Basin Formation

These actively evolving basins range from foreland basins partly stuffed with marine sediments and with existing water depths of 1500–3000 m [for example, the Seram Trough (Audley-Charles & Carter 1977) and the Timor-Tanimbar Trough (Audley-Charles 1986)] to fully stuffed foreland basins with thick carbonate sediments at the base giving way upward to nonmarine, detrital sediments [as in the Papuan Basin (Pigram et al 1989)]. They appear to be forming, at least in part, in response to the crustal loading effect of the nappes in the orogenic hinterland (Jordan 1981). Price & Audley-Charles (1987) have suggested that where the collision zone faults have locked, as in Timor, the lithosphere is buckling in response to continuing plate convergence.

Reversal of Subduction Polarity

The development of subduction reversal after arc-continent collision has been discussed by Hamilton (1979). Along the New Guinea Trench (Figure 1), west of 144°E, seismicity defines a southward-dipping Wadati-Benioff zone in which focal mechanisms indicate oblique subduction, which may represent reversal in subduction polarity, according to Cooper & Taylor (1987). Silver et al (1983), discussing the indications of back-arc thrusting related to arc-continent collision in the Banda arc, concluded that this back-arc thrusting could lead to subduction reversal. Price & Audley-Charles (1987), from their modeling of the Banda collision zone, proposed that this reversal is active.

Influence of Crustal Heterogeneities

The Ontong-Java oceanic plateau, about 2400 km long and 1000 km wide and having oceanic-type crust five times normal thickness (Nur & Ben-Avraham 1982), is in collision with the north Solomon Trench (Figure 1).

According to Kroenke (1984), this collision began in the early Miocene (22–20 Ma), terminating subduction here. Hall & Nichols (1990) have shown how thickened crust of oceanic plateaus can deflect the path of propagating trenches and strike-slip faults and lead to the relatively passive accretion of oceanic plateaus at evolving plate boundaries. The eastward propagation of the Late Cretaceous Sunda Trench (Daly et al 1987) during the Miocene, giving rise to the Banda arc, appears to have been associated with the presence of crustal heterogeneities in the form of continental fragments with amalgamated Paleogene volcanic terranes (Harris 1989).

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