Tectonics of the New Guinea Region

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Abstract

The New Guinea region evolved within the obliquely and rapidly converging Australian and Pacific plate boundary zone. It is arguably one of the most tectonically complex regions of the world, and its geodynamic evolution involved microplate formation and rotation, lithospheric rupture to form ocean basins, arc-continent collision, subduction polarity reversal, collisional orogenesis, ophiolite obduction, and exhumation of (ultra)high-pressure metamorphic rocks. We describe the major onshore and offshore tectonic and geologic components, including plate boundaries, seismicity, faults, and magmatism, and we integrate these with emerging ideas about mantle dynamics to evaluate the Cenozoic tectonic evolution of New Guinea. Future research opportunities to resolve the mantle structure beneath New Guinea will enable mantle dynamics to be linked to lithospheric and surface processes. Virtually all plate tectonic and mantle processes have been active in the New Guinea region throughout the Cenozoic, and, as such, its tectonic evolution has global significance.

INTRODUCTION

The island of New Guinea, comprising the Independent State of Papua New Guinea in the east and the Republic of Indonesia in the west, is evolving within the obliquely and rapidly converging Australian (AUS) and Pacific (PAC) plate boundary zone (**Figure 1**). It is arguably one of the most tectonically complex regions of the world and is in the vanguard for understanding the full gamut

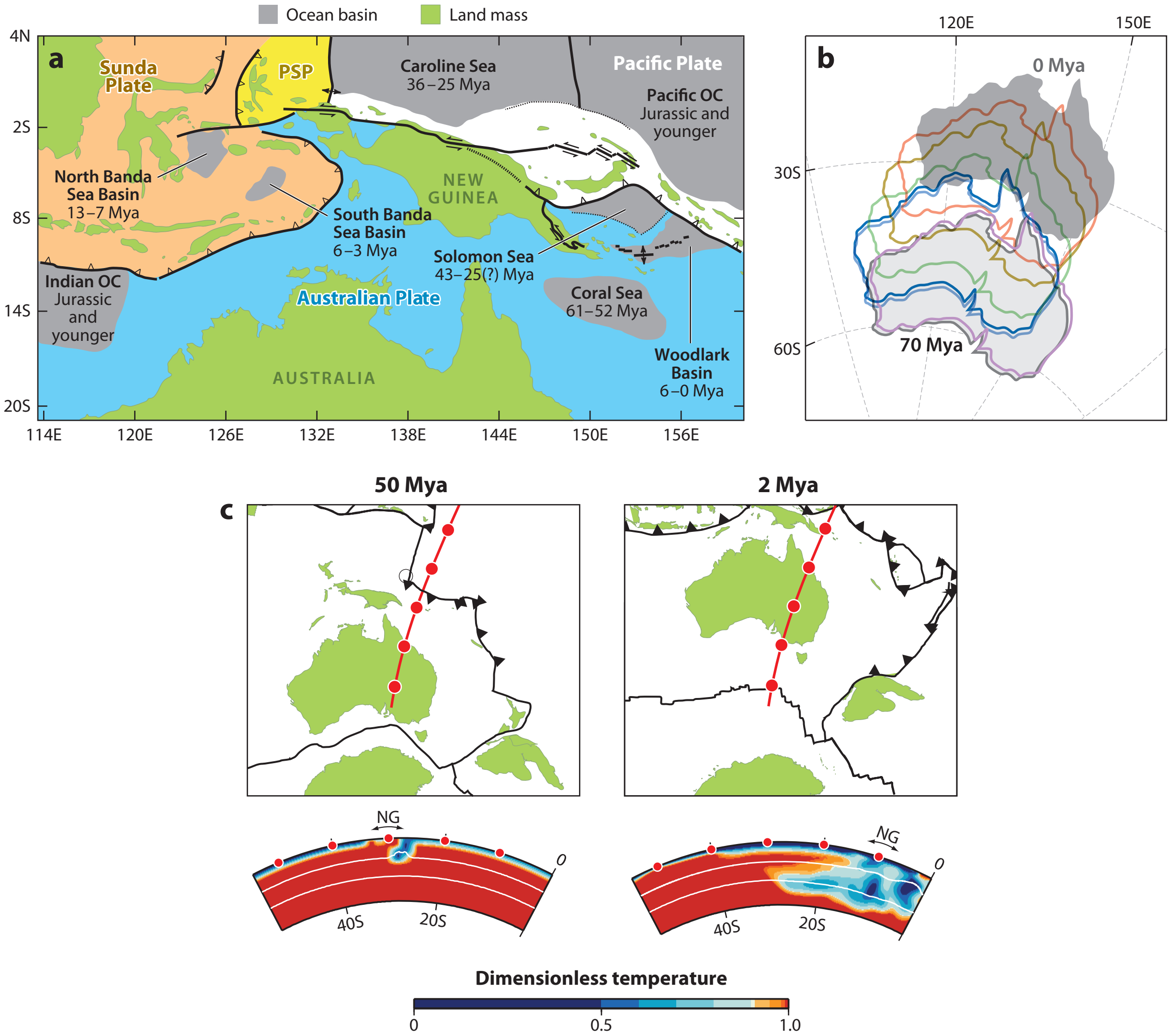


Figure 1

(*a*) Primary tectonic plates of the New Guinea region, with land masses shown in green: Australian plate, Pacific plate, Sunda plate (*orange*), and Philippine Sea plate (PSP; *yellow*). Ocean basins are shown in gray, with approximate ages of oceanic lithosphere indicated. Abbreviation: OC, oceanic crust. (*b*) Northward motion of Australia in an absolute reference frame since 70 million years ago (Mya) (modified from Heine et al. 2010). (*c*) Plate reconstructions and paleothermal models at 50 and 2 Mya (modified from DiCaprio et al. 2011). Models show Australia plowing north over Melanesian subducted slabs. Position of New Guinea (NG) is projected into the plane of section. White lines in cross sections represent the 410- and 660-km phase changes.

of tectonic processes on Earth including the opening and closing of ocean basins (i.e., the Wilson cycle), terrane accretion, ophiolite obduction, subduction reversal, and ultrahigh-pressure (UHP) rock exhumation.

New Guinea has abundant mineral and petroleum resources and a long history of mining and exploration, notably for gold, silver, and copper (Williamson & Hancock 2005). Some successor basins that overlap terrane boundaries and sutures are prolific producers of oil and gas (e.g., Bintuni; Doust & Noble 2008).

A vast amount of literature dating from 1850 addresses the geology of New Guinea (for a comprehensive bibliography, see **http://www.vangorselslist.com/new_guinea.html**). Reviews of the tectonic evolution of the New Guinea region include Hamilton (1979), Pigram & Davies (1987), Audley-Charles (1991), Hill & Hall (2003), and Cloos et al. (2005), as well as those that discuss reconstructions of surrounding ocean basins (Schellart et al. 2006, Gaina & Müller 2007) and paleogeographic and plate reconstructions (e.g., Hall 2002, Spakman & Hall 2010). Considerable controversy exists regarding the Cenozoic tectonic evolution of New Guinea, in part due to its remoteness and to the complex and rapidly changing tectonic settings. In this review, we integrate results from onshore and offshore studies with emerging data sets that provide clues as to how mantle dynamics may have impacted the tectonic evolution of the New Guinea region.

PRESENT-DAY TECTONIC SETTING OF THE NEW GUINEA REGION

Cenozoic northward movement of the AUS plate through almost 30° of latitude (Müller et al. 2008) (**Figure 1*b***) and oblique collision with the PAC plate result in a complex plate boundary zone within which New Guinea evolved (Johnson & Molnar 1972, Hall & Spakman 2002, Gaina et al. 2007) (**Figure 1*a***). The Cenozoic evolution of northwestern New Guinea also involved interaction with the Sunda plate (Bird 2003), resulting in the accretion of lithospheric fragments.

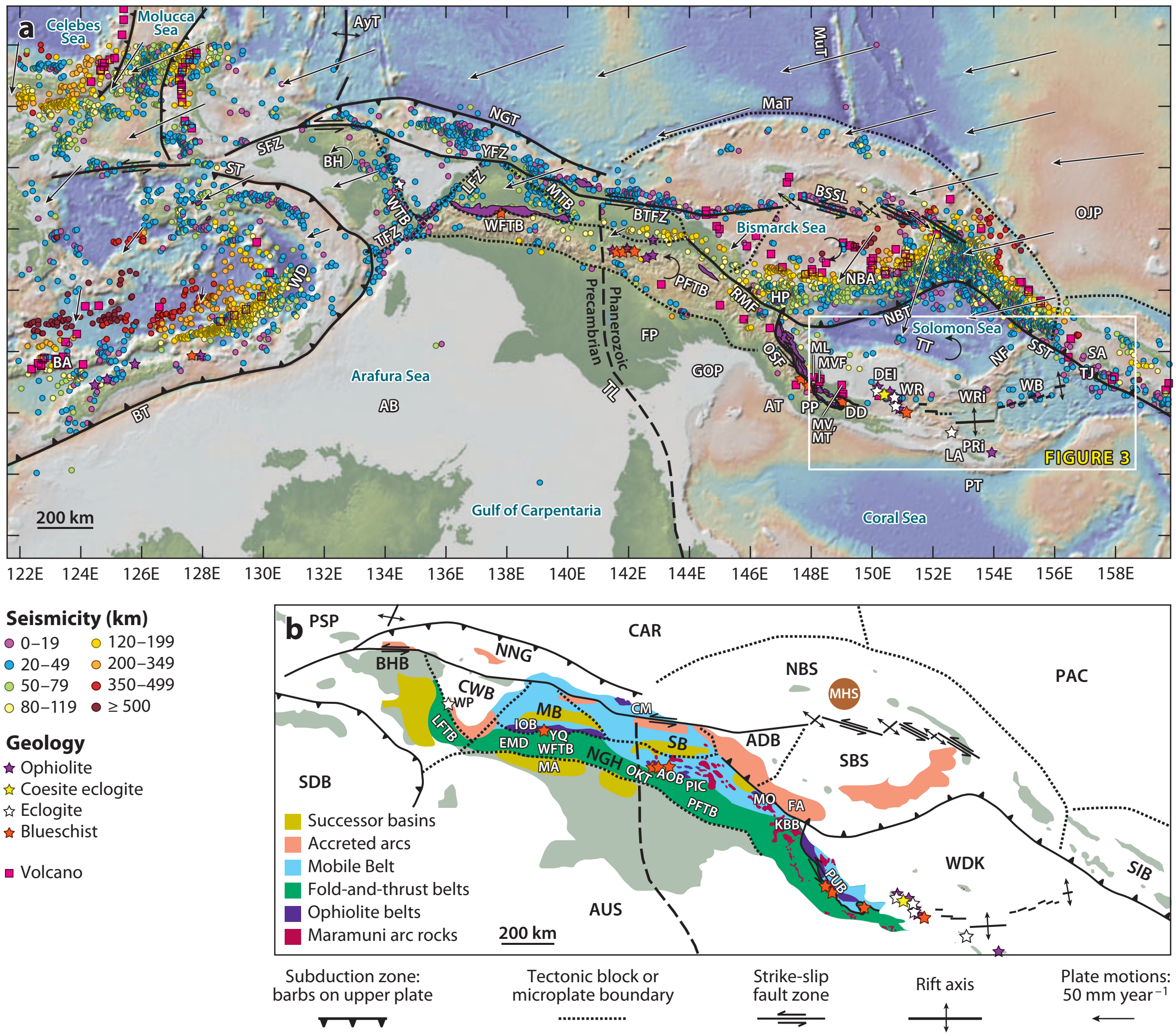
The AUS-PAC plate boundary zone is dominated by west-southwest motion of the PAC plate relative to the AUS plate, at ~110 mm year^{-1} (DeMets et al. 1994) (**Figure 2*a***), with a convergent component of ~70 mm year^{-1} across the New Guinea region (Tregoning & Gorbatov 2004). The AUS-PAC plate boundary zone has considerable along-strike variation (e.g., Hall & Spakman 2002; Wallace et al. 2004, 2005), is highly dynamic (Hall 2002, Baldwin et al. 2004, Bailly et al. 2009), and is seismically active (**http://neic.usgs.gov;** see also **Supplemental Figure 1**, accessible from the **Supplemental Materials link** on the Annual Reviews home page at **http://www.annualreviews.org/**). Some subducting slabs have extreme curvature in map view (Banda, Celebes, New Britain, South Solomons), and others have changes in slab dip over relatively short distances (New Guinea Trench). Other seismically active regions are associated with seafloor spreading (Woodlark Basin, Bismarck Sea), rifting (Woodlark Rift), collisional orogenesis (New Guinea highlands), and strike-slip faulting (e.g., Sorong fault zone).

Pacific and Australian Major Plate Boundaries in the New Guinea Region

The PAC oceanic lithospheric plate is long-lived [>180 million years old (Ma old)] and the largest plate on Earth. On its southwestern boundary the Ontong Java Plateau (OJP), the largest and thickest (33 km) oceanic plateau (e.g., Taylor 2006), collided with the Solomon arc, resulting in subduction reversal in the Late Miocene (e.g., Mann & Taira 2004). Seismicity indicates the presence of mutually inward-dipping Benioff zones, revealing that although subduction reversal has occurred, active subduction of the leading edge of the OJP is still ongoing (Miura et al. 2002). The nature of boundaries between the PAC plate and the Bismarck microplates is less certain (**Figure 2**). GPS constraints (Wallace et al. 2004) indicate relatively minor convergence between the PAC plate and the North Bismarck (NBS) microplate at 7–10 mm year^{-1}. Further west, where the PAC plate abuts the Caroline microplate (Gaina & Müller 2007), their similar velocities (McCaffrey 1996) plus the lack of seismicity at the Caroline-PAC boundary suggest that the Caroline microplate is moving with the PAC plate (Bird 2003).

The AUS plate includes the Australian continent, rifted continental fragments (e.g., Lord Howe Rise, Norfolk Rise), New Zealand west of the Alpine fault, and surrounding ocean basins (e.g., Coral Sea). The AUS plate is moving north-northeast relative to a fixed PAC plate (DeMets et al. 1994, Bird 2003) (**Figure 1*b***). Geodetic, seismic, and geological data indicate that the northern margin of the AUS plate in southern New Guinea is actively deforming (Tregoning 2003). Stable areas of the northern AUS plate include the wide continental shelf that links northern Australia with New Guinea and extends east to the Coral Sea.

The boundary between the AUS plate and surrounding (micro)plates varies considerably along strike. Southeast of the AUS-Woodlark (WDK)-PAC triple junction (Chadwick et al. 2009) (**Figure 3**), the AUS plate subducts beneath the Solomon Islands at the South Solomons



Trench. Southwest of the AUS-WDK-PAC triple junction, the AUS-WDK plate boundary is the Woodlark Basin spreading center (**Figure 3**). West of the Woodlark Basin seafloor spreading rift tip, the AUS-WDK plate boundary transitions from divergence to convergence over 500 km, from the Woodlark Rift to the Papuan peninsula, then along the Owen-Stanley fault zone to link up with the Ramu-Markham fault. Deformation along the Owen-Stanley fault zone is dominated by extension near the tip of the Papuan peninsula, oblique-sinistral slip (where the Dayman Dome is currently exhuming; Daczko et al. 2011), sinistral strike-slip faulting, and finally ~20 mm $year^{-1}$ of convergence as it nears the intersection with the Ramu-Markham fault (Wallace et al. 2004).

In central New Guinea, the AUS plate obliquely converges with the South Bismarck (SBS), NBS, and Caroline microplates (**Figure 2**). In northwestern New Guinea, near 134°E, the northern margin of the stable AUS plate meets the tightly curved Banda arc, where subduction of a (now) folded slab and associated rollback may have engulfed an inferred Jurassic embayment of oceanic lithosphere within the AUS plate (Spakman & Hall 2010).

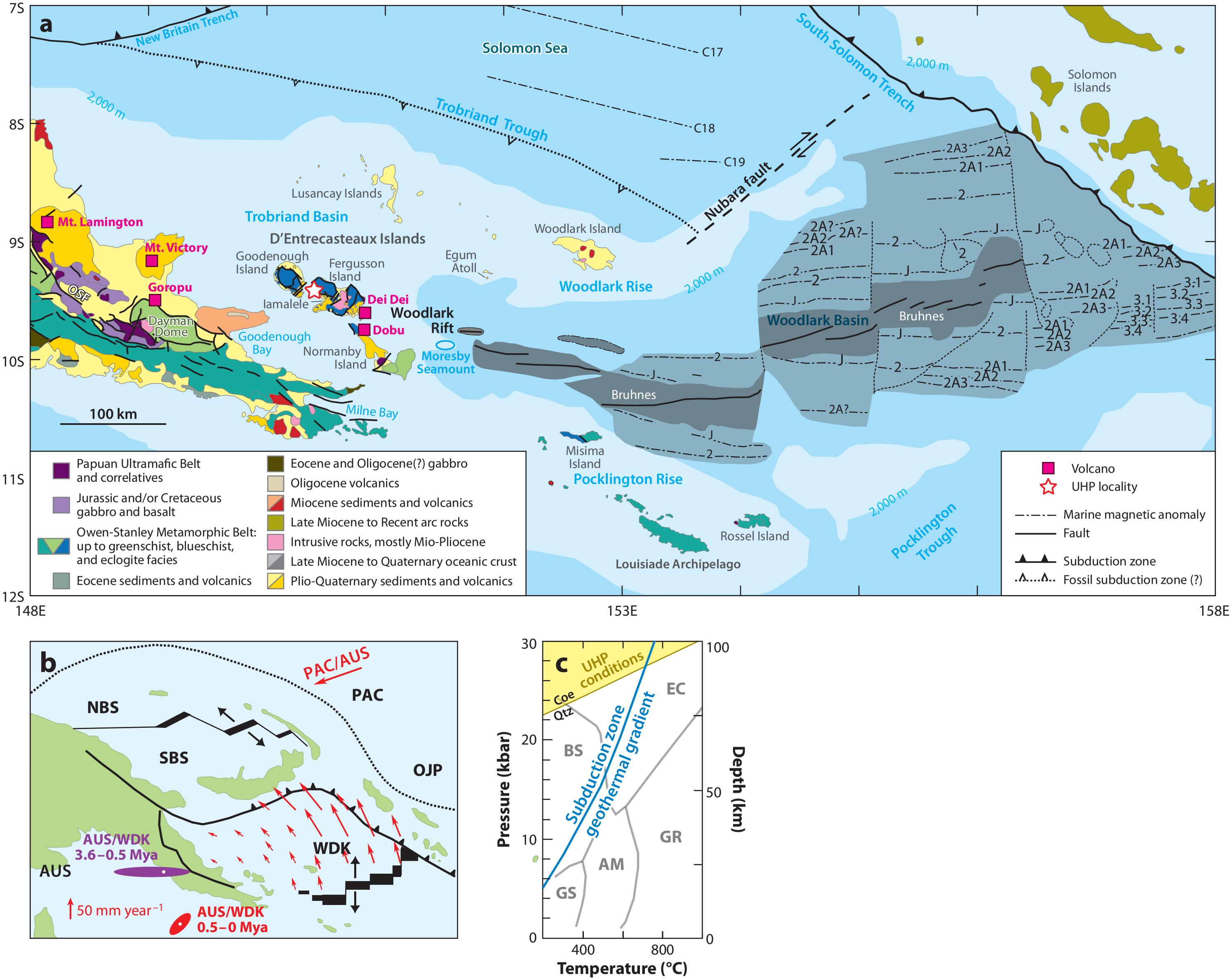
Microplates in the New Guinea Region and Their Relevant Boundaries

The number and extent of microplates that formed, rotated, and evolved (perhaps also meeting their demise) within the larger AUS-PAC plate boundary zone are controversial. In map view, the AUS-PAC zone of deformation resembles a large left-lateral sigma clast, within which microplates evolved. The age of oceanic lithosphere in microplates surrounding New Guinea ranges from Eocene to present (**Figure 1*a***). Aspects of microplate evolution are discussed below, beginning with the WDK microplate and proceeding in an anticlockwise direction.

Woodlark microplate. The WDK microplate is bounded to the north by the New Britain Trench and the South Solomon Trench and to the south by the active Woodlark seafloor

Figure 2

Tectonic maps of the New Guinea region. (*a*) Seismicity, volcanoes, and plate motion vectors. Plate motion vectors relative to the Australian plate are surface velocity models based on GPS data, fault slip rates, and earthquake focal mechanisms (UNAVCO, **http://jules.unavco.org/Voyager/Earth**). Earthquake data are sourced from the International Seismological Center EHB Bulletin (**http://www.isc.ac.uk**); data represent events from January 1994 through January 2009 with constrained focal depths. Background image is generated from **http://www.geomapapp.org**. Abbreviations: AB, Arafura Basin; AT, Aure Trough; AyT, Ayu Trough; BA, Banda arc; BSSL, Bismarck Sea seismic lineation; BH, Bird's Head; BT, Banda Trench; BTFZ, Bewani-Torricelli fault zone; DD, Dayman Dome; DEI, D'Entrecasteaux Islands; FP, Fly Platform; GOP, Gulf of Papua; HP, Huon peninsula; LA, Louisiade Archipelago; LFZ, Lowlands fault zone; MaT, Manus Trench; ML, Mt. Lamington; MT, Mt. Trafalgar; MuT, Mussau Trough; MV, Mt. Victory; MTB, Mamberamo thrust belt; MVF, Managalase Plateau volcanic field; NBT, New Britain Trench; NBA, New Britain arc; NF, Nubara fault; NGT, New Guinea Trench; OJP, Ontong Java Plateau; OSF, Owen Stanley fault zone; PFTB, Papuan fold-and-thrust belt; PP, Papuan peninsula; PRi, Pocklington Rise; PT, Pocklington Trough; RMF, Ramu-Markham fault; SST, South Solomons Trench; SA, Solomon arc; SFZ, Sorong fault zone; ST, Seram Trench; TFZ, Tarera-Aiduna fault zone; TJ, AUS-WDK-PAC triple junction; TL, Tasman line; TT, Trobriand Trough; WD, Weber Deep; WB, Woodlark Basin; WFTB, Western (Irian) fold-and-thrust belt; WR, Woodlark Rift; WRi, Woodlark Rise; WTB, Weyland thrust; YFZ, Yapen fault zone. White box indicates the location shown in **Figure 3**. (*b*) Map of plates, microplates, and tectonic blocks and elements of the New Guinea region. Tectonic elements modified after Hill & Hall (2003). Abbreviations: ADB, Adelbert block; AOB, April ultramafics; AUS, Australian plate; BHB, Bird's Head block; CM, Cyclops Mountains; CWB, Cendrawasih block; CAR, Caroline microplate; EMD, Ertsberg Mining District; FA, Finisterre arc; IOB, Irian ophiolite belt; KBB, Kubor & Bena blocks (including Bena Bena terrane); LFTB, Lengguru fold-and-thrust belt; MA, Mapenduma anticline; MB, Mamberamo Basin block; MO, Marum ophiolite belt; MHS, Manus hotspot; NBS, North Bismarck plate; NGH, New Guinea highlands block; NNG, Northern New Guinea block; OKT, Ok Tedi mining district; PAC, Pacific plate; PIC, Porgera intrusive complex; PSP, Philippine Sea plate; PUB, Papuan Ultramafic Belt ophiolite; SB, Sepik Basin block; SDB, Sunda block; SBS, South Bismarck plate; SIB, Solomon Islands block; WP, Wandamen peninsula; WDK, Woodlark microplate; YQ, Yeleme quarries.


a
7S
8S
9S
10S
11S
12S
148E
153E
158E
New Britain Trench
Solomon Sea
C17
C18
C19
South Solomon Trench
Solomon Islands
Trobriand Trough
Nubara fault
2,000 m
Lusancay Islands
Trobriand Basin
D'Entrecasteaux Islands
Goodenough Island
Fergusson Island
Iamalele
Egum Atoll
Woodlark Island
Woodlark Rise
Mt. Lamington
Mt. Victory
Goropu
OSF
Dayman Dome
Dei Dei
Dobu
Woodlark Rift
Goodenough Bay
Normanby Island
Moresby Seamount
Woodlark Basin
Bruhnes
Milne Bay
100 km
Misima Island
Pocklington Rise
Rossel Island
Louisiade Archipelago
Pocklington Trough
2A1
2A2
2A3
2A?
2
J
3.1
3.2
3.3
3.4
Papuan Ultramafic Belt and correlatives
Jurassic and/or Cretaceous gabbro and basalt
Owen-Stanley Metamorphic Belt: up to greenschist, blueschist, and eclogite facies
Eocene sediments and volcanics
Eocene and Oligocene(?) gabbro
Oligocene volcanics
Miocene sediments and volcanics
Late Miocene to Recent arc rocks
Intrusive rocks, mostly Mio-Pliocene
Late Miocene to Quaternary oceanic crust
Plio-Quaternary sediments and volcanics
Volcano
UHP locality
Marine magnetic anomaly
Fault
Subduction zone
Fossil subduction zone (?)
b
NBS
SBS
PAC/AUS
PAC
OJP
WDK
AUS
AUS/WDK 3.6–0.5 Mya
50 mm year^{-1}
AUS/WDK 0.5–0 Mya
c
UHP conditions
Coe
Qtz
EC
BS
GR
AM
GS
Subduction zone geothermal gradient
Pressure (kbar)
Depth (km)
Temperature (°C)
0
10
20
30
50
100
400
800

spreading center (**Figure 3**). The western end of the spreading center transitions into the most rapidly extending (~20–40 mm year^{-1} at 151.5°E) active rift system on Earth (e.g., Abers 2001). Since ~6 million years ago (Mya), the Woodlark Basin spreading center rift tip has propagated >500 km westward at ~14 cm year^{-1} via a stepwise process with periods of spreading center nucleation, propagation, and stalling, resulting in the separation of the once contiguous Woodlark Rise (northern rifted margin) and Pocklington Rise (southern rifted margin) (Taylor et al. 1995, 1999).

Spreading from >3.5 to 0.5 Mya was faster (4.2° per Ma) and oriented more north-south than it has been since 0.5 Mya, when spreading slowed (2.4° per Ma) and became more northwest-southeast (Taylor et al. 1999). At the seafloor spreading rift tip located near the Moresby Seamount, extension is 20 mm year^{-1} (Wallace et al. 2004). Extension across the rift is accommodated on active low-angle (25–30°) (Abers & Roecker 1991, Abers et al. 1997, Abers 2001) and high-angle normal faults (Kington & Goodliffe 2008). West of the Moresby Seamount, metamorphic core complexes of the D'Entrecasteaux Islands lie within the active Woodlark Rift (e.g., Davies & Warren 1988, Hill et al. 1992, Baldwin et al. 1993, Little et al. 2007).

Seafloor spreading in the Woodlark Basin and rifting in the Woodlark Rift are apparently driven by subduction of the Solomon Sea to the north to form the New Britain arc (Weissel et al. 1982, Hall 2001). Some workers have classified the Solomon Sea as its own microplate (e.g., Benes et al. 1994) either in lieu of the WDK microplate or in addition to it. Magnetic lineations (41.5–35 Mya) are present in the Solomon Sea, although spreading likely initiated ~43–41 Mya and possibly continued until ~25 Mya (Joshima et al. 1987, Gaina & Müller 2007). No ridge is observed, and only a portion of the southern flank of the basin is preserved, with the remainder lost to northward subduction.

Some studies have argued for two oppositely dipping subduction zones, north-dipping at the New Britain Trench and south-dipping at the Trobriand Trough (Cooper & Taylor 1987, Pegler et al. 1995). A separate Trobriand microplate (Kington & Goodliffe 2008) has also been proposed with the east-southeast-trending Trobriand Trough, marking recently active southward subduction. However, seismicity at the Trobriand Trough lacks an organized (shallow-to-deep) pattern indicative of southward subduction. Seismic reflection images of the Trobriand Trough show weakly deformed sediment and normal faulting, allowing convergence of no more than a few millimeters per year (Davies & Jaques 1984, Lock et al. 1987, Kirchoff-Stein 1992). Overall, available geophysical data suggest that subduction at the Trobriand Trough is presently inactive (e.g., Abers et al. 2002, Hall & Spakman 2002, Wallace et al. 2004), although it may have been active in the Miocene (see below and see also sidebar, Outstanding Questions).

South and North Bismarck microplates. The Bismarck Sea is a back-arc basin with respect to the New Britain arc. It is divided into the North Bismarck (NBS) and South Bismarck (SBS) microplates by the active Bismarck Sea seismic lineation, a zone of left-lateral transform faults and spreading segments (Taylor 1979, Tregoning et al. 1999) (**Figure 2**). The Manus Trench,

Figure 3

(*a*) Tectonic and geological map of the Woodlark Rift region. Geology modified from Davies (1971, 1980b); magnetic anomalies from Taylor et al. (1995) and Gaina & Müller (2007). (*b*) The tectonic setting of the Woodlark Rift. Pole of rotation for 3.6–0.5 million years ago (Mya) (Taylor et al. 1999) and present-day pole of rotation (Wallace et al. 2004) are shown with error ellipses. (*c*) Location of metamorphic facies in pressure/temperature space. Abbreviations in panels *a* and *b* are as in **Figure 2**. Abbreviations in panel *c*: AM, amphibolite facies; BS, blueschist facies; Coe, coesite; EC, eclogite facies; GR, granulite facies; GS, greenschist facies; Qtz, quartz; UHP, ultrahigh-pressure.

OUTSTANDING QUESTIONS

Globally Significant Questions

1. How did the geoid high in the New Guinea region form, and what is its significance with respect to the tectonic evolution?
2. What does the slab accumulation under New Guinea look like, and how has it changed during the Cenozoic as the AUS plate moved northward?
3. Has mantle flow at the western edge of the Jason LLSVP impacted the New Guinea region, and if so, how?
4. How has Pliocene to Recent surface uplift of the New Guinea landmass impacted global ocean circulation patterns and climate?

Regionally Significant Questions

1. How has the lithosphere evolved as the leading edge of the AUS plate plows up subduction complexes, obducts ophiolites, and collides with island arc(s) during its north-northeast passage across a complexly structured mantle? How has this impacted mantle flow beneath the southwest Pacific? What was the dynamic topographic response during this north-northeast passage?
2. How is deformation partitioned during AUS-PAC oblique convergence? What are the relative roles of extension versus strike-slip faulting? What is causing the apparent southward-younging trend of deformation and magmatism?
3. How many island arcs existed north of the AUS plate during the Cenozoic, how did they form, what was their polarity, and what was their accretion history? How are tectonic events in the New Guinea region linked in space and time?

accommodating mainly sinistral translation, defines the boundary between the NBS microplate and the Caroline/PAC plates. Magnetic anomalies in the Bismarck Sea indicate rapid asymmetric spreading since 3.5 Mya (Taylor 1979). The SBS microplate encompasses the New Britain arc, with its southern boundary the New Britain Trench and its northern boundary the Bismarck Sea seismic lineation (e.g., Taylor 1979, Gaina & Müller 2007). The southwest boundary of the SBS microplate with the AUS plate is the Ramu-Markham fault, formed when the Finisterre arc collided with northern New Guinea in the Late Miocene (e.g., Abbott et al. 1994, Hill & Raza 1999). The SBS microplate is rotating rapidly clockwise (~9° Ma^{-1}), likely a result of the arc-continent collision, whereas the NBS microplate is rotating slowly anticlockwise (0.3–1.25° Ma^{-1}) (Wallace et al. 2004, 2005).

Caroline microplate. North-central New Guinea is bounded by the Caroline microplate (Weissel & Anderson 1978, Bird 2003). Magnetic anomalies (36–25 Mya) indicate formation in the Late Eocene–Oligocene with mainly north-south spreading (Gaina & Müller 2007). Hall (2002) considers the Caroline microplate as a back-arc basin associated with westward subduction of the PAC plate under the Philippine Sea plate. Gaina & Müller (2007) model the Caroline microplate as a back-arc basin north of a north-dipping subducting AUS slab, but east of the Philippine Sea plate, with the arc later accreting to northern New Guinea during Late Oligocene–Miocene arc-continent collision.

The Caroline microplate is bounded to the southwest by the Ayu Trough, a spreading center probably of Late Oligocene (~25 Mya) (Fujiwara et al. 1995) or Miocene age (Weissel & Anderson

1978), likely formed by anticlockwise rotation (e.g., Hegarty & Weissel 1988). Whether the New Guinea Trench represents active Caroline microplate subduction beneath northern New Guinea is debated. Using tomography, Tregoning & Gorbatov (2004) argue for a 650-km-long slab to depths of 300 km with a dip angle decreasing from $\sim$30° to $\sim$10°, whereas Hall & Spakman (2002, 2003) found no evidence for a subducting slab at the New Guinea Trench. Cloos et al. (2005) agree with Kroenke (1984) that the New Guinea Trench is a relict of earlier subduction and that only the eastern part of the trench has been reactivated owing to arc/forearc-continent collision (i.e., collision of the Melanesian arc discussed below). On the basis of GPS constraints, up to 40 mm year^{-1} of AUS-PAC convergence could be accommodated by subduction of the PAC plate at the New Guinea Trench (Puntodewo et al. 1994, McCaffrey 1996, Stevens et al. 2002).

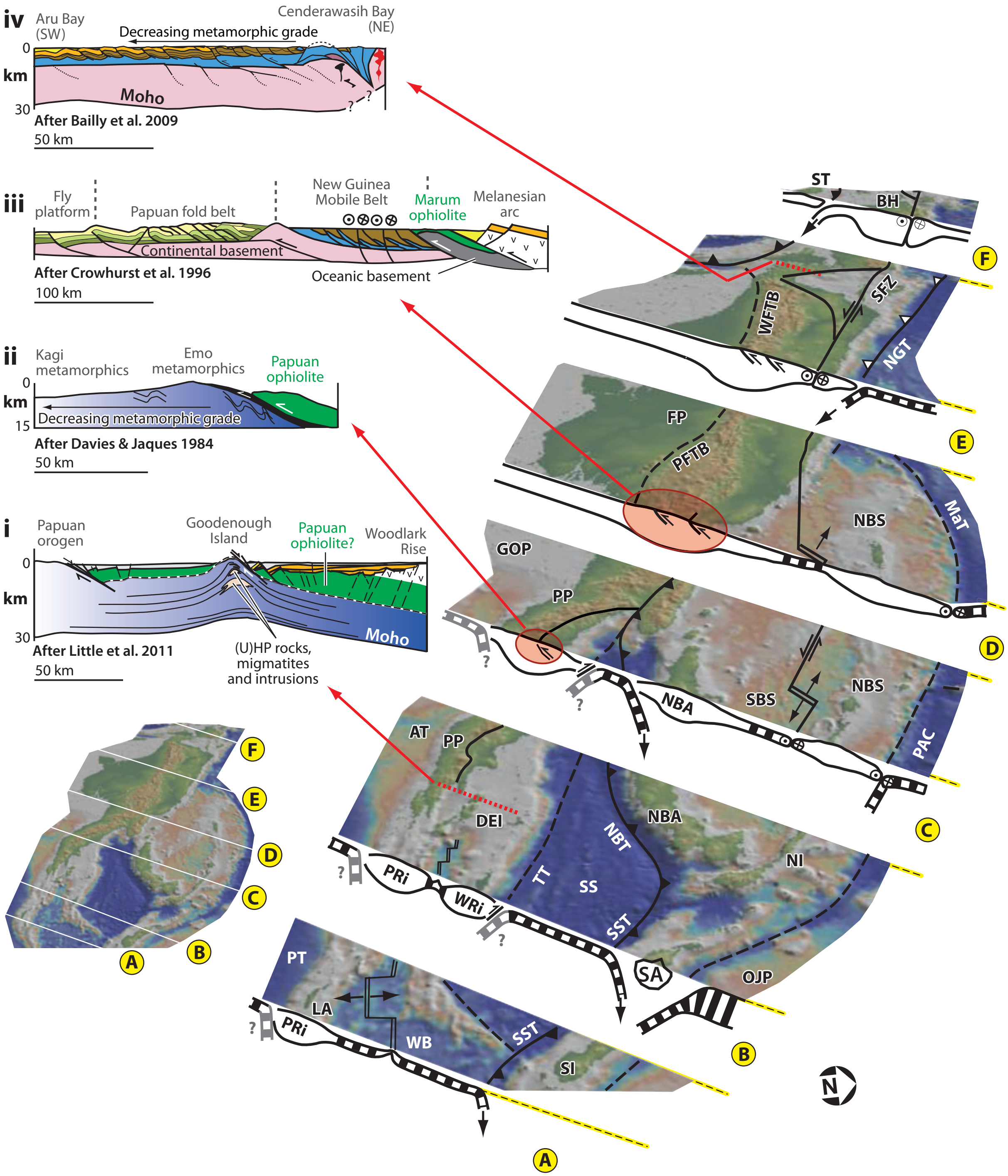
TECTONICS OF NEW GUINEA

During the Cenozoic, New Guinea occupied the leading northern edge of the AUS plate as it plowed northward, progressively creating and colliding with island arcs and microplates. Remnants of accreted arcs, subduction complexes, ophiolites, and microplates preserve a record of multiple episodes of subduction/obduction and rock and surface uplift. Schematic north-south geologic cross sections of New Guinea in **Figure 4** illustrate the distribution of continental margin sediments deformed in fold-and-thrust belts, variably metamorphosed Jurassic and Cretaceous sediments (i.e., the products of subduction-zone metamorphism), obducted fragments of oceanic lithosphere, and volcanic and plutonic rocks.

Active Deformation

Major active faults include, from east to west: the Owen-Stanley fault zone, the Ramu-Markham fault, thrust and strike-slip faults in central New Guinea, and strike-slip faults along the northern New Guinea margin (e.g., the Yapen and Sorong fault zones). The Owen-Stanley fault zone (Davies & Smith 1971) separates the WDK plate from the New Guinea highlands block, linking via a series of transfer faults from the Woodlark spreading center rift tip at the Moresby Seamount (e.g., Little et al. 2007), to the Papuan peninsula (**Figures 2** and **3**). The Owen-Stanley fault zone is a reactivated megathrust that initially accommodated southwestward obduction of the Papuan ophiolite during arc-continent collision in the Paleogene, over the Owen-Stanley metamorphics, which were scraped off the orogenic wedge of the AUS plate (Davies 1980a, Davies & Jaques 1984). Little et al. (2011) defined a D'Entrecasteaux fault zone that is correlative with, and includes, the Owen-Stanley fault zone as the former thrust fault along which ophiolite obduction occurred. This fault has since been reactivated to form active normal faults that exhume the Dayman Dome and active normal faults at the base of the upper plate of the D'Entrecasteaux Islands core complexes. To the west, other reactivated faults south of the obducted April ophiolite (Lagaip fault) and the Irian ophiolite (Derewo fault, which defines a 200-km straight valley) likely accommodated earlier thrusting and obduction (Hill & Hall 2003) and are now largely accommodating left-lateral slip associated with oblique collision (Cloos et al. 2005).

The Ramu-Markham thrust fault accommodates motion between the colliding Finisterre arc (on the SBS plate) and the New Guinea highlands block (AUS plate) (Pegler et al. 1995, Davies et al. 1997) as the Finisterre arc is accreted onto northern New Guinea. Deformation (slip) along the Ramu-Markham fault is greater in the west than in the east, indicating oblique collision of the Melanesian arc with New Guinea (Wallace et al. 2004). Earthquakes and seismic velocities reveal a well-defined north-dipping slab beneath the Finisterre arc to depths of 250 km that continues along


iv
Aru Bay (SW)
Cenderawasih Bay (NE)
Decreasing metamorphic grade
0
km
30
Moho
After Bailly et al. 2009
50 km
iii
Fly platform
Papuan fold belt
New Guinea Mobile Belt
Marum ophiolite
Melanesian arc
Continental basement
Oceanic basement
After Crowhurst et al. 1996
100 km
ii
Kagi metamorphics
Emo metamorphics
Papuan ophiolite
0
km
15
Decreasing metamorphic grade
After Davies & Jaques 1984
50 km
i
Papuan orogen
Goodenough Island
Papuan ophiolite?
Woodlark Rise
0
km
30
Moho
After Little et al. 2011
50 km
(U)HP rocks, migmatites and intrusions
ST
BH
F
WFTB
SFZ
NGT
E
FP
PFTB
NBS
MaT
GOP
PP
D
NBA
SBS
NBS
PAC
AT
PP
DEI
NBA
C
NBT
NI
TT
SS
SST
PRi
WRi
SA
OJP
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LA
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strike to New Britain, where earthquakes reach depths of 600 km (Abers & Roecker 1991). Beneath the accreted Finisterre arc, seismic activity defines an inverted-U-shape pattern, representing the steeply dipping northern slab that may be mechanically decoupled from the overlying lithosphere at ~100 km, and also defines a relict south-dipping slab (Pegler et al. 1995, Woodhead et al. 2010).

Sinistral strike-slip faults in the northern part of New Guinea (e.g., Cloos et al. 2005), taking up lateral motion due to oblique arc collision, include the Bewani-Torricelli fault zone that links to transform faults offshore, which, in turn, link to seafloor spreading segments in the Bismarck Sea. To the west, the Bewani-Torricelli fault zone links to the Yapen left-lateral fault system through the Mamberamo thrust belt, a region of low elevation characterized by active mud volcanoes and thrust-dominated deformation. Further west, the Yapen connects to the Sorong fault zone (**Figure 2**).

Seismic studies demonstrate partitioning of deformation in the New Guinea highlands into thrust and strike-slip components (Abers & McCaffrey 1988). However, modeling of GPS data suggests that the New Guinea highlands fits a rigid block model, consistent with ~1.6° Ma^{-1} of anticlockwise rotation and up to 15 mm $year^{-1}$ of convergence in the highlands and Mamberamo thrust belt (Puntodewo et al. 1994, Wallace et al. 2004).

The New Guinea highlands block is bound to the south by the thrust front of the fold-and-thrust belts, which is roughly the southern limit of collisional orogenesis as defined by both seismicity and topography. The Mamberamo Basin block lies east of the Lowlands fault zone, a zone of active sinistral strike-slip and reverse faulting. The southern boundary of the Mamberamo Basin block corresponds to the tectonic contact between the Irian ophiolite belt and the western fold-and-thrust belt of the New Guinea highlands block. East of the Mamberamo Basin block lies the Sepik Basin block, with the boundary delineated by shallow seismicity within the Mamberamo thrust zone. The southern boundary of the Sepik Basin block is roughly associated with a belt of intermediate-depth seismicity continuous with that of the New Britain arc. On the basis of tomography, the intermediate-depth seismicity in these regions may be associated with southward subduction at the New Guinea Trench (Tregoning & Gorbatov 2004).

The Bird's Head block is bound to south by the left-lateral Tarera-Aiduna fault zone and to the east by the Weyland thrust belt. East of the Bird's Head is the Cendrawasih block, the eastern boundary of which is the Lowlands fault zone. GPS data (Stevens et al. 2002) indicate that the Bird's Head is a relatively rigid block rotating anticlockwise (~1.8° Ma^{-1}) relative to the AUS plate, consistent with faster convergence rates at the Seram Trench, where there is subduction

Figure 4

Oblique block diagram of New Guinea from the northeast with schematic cross sections showing the present-day plate tectonic setting. Digital elevation model was generated from **http://www.geomapapp.org**. Oceanic crust in tectonic cross sections is shown by thick black-and-white hatched lines, with arrows indicating active subduction; thick gray-and-white hatched lines indicate uncertain former subduction. Continental crust, transitional continental crust, and arc-related crust are shown without pattern. Representative geologic cross sections across parts of slices C and D are marked with transparent red ovals and within slices B and E are shown by dotted lines. (*i*) Cross section of the Papuan peninsula and D'Entrecasteaux Islands modified from Little et al. (2011), showing the obducted ophiolite belt due to collision of the Australian (AUS) plate with an arc in the Paleogene, with later Pliocene extension and exhumation to form the D'Entrecasteaux Islands. (*ii*) Cross section of the Papuan peninsula after Davies & Jaques (1984) shows the Papuan ophiolite thrust over metamorphic rocks of AUS margin affinity. (*iii*) Across the Papuan mainland, the cross section after Crowhurst et al. (1996) shows the obducted Marum ophiolite and complex folding and thrusting due to collision of the Melanesian arc (the Adelbert, Finisterre, and Huon blocks) in the Late Miocene to recent. (*iv*) Across the Bird's Head, the cross section after Bailly et al. (2009) illustrates deformation in the Lengguru fold-and-thrust belt as a result of Late Miocene–Early Pliocene northeast-southwest shortening, followed by Late Pliocene–Quaternary extension. Abbreviations as in **Figure 2**, in addition to NI, New Ireland; SI, Solomon Islands; SS, Solomon Sea; (U)HP, (ultra)high-pressure.

of continental crust. Stevens et al.'s (2002) GPS data suggest that the Cendrawasih block is part of the Bird's Head block, moving 75–80 mm year^{-1} relative to the AUS plate along a trajectory similar to that of the PAC plate. Much of that west-southwest motion is accommodated by left-lateral slip and fault block rotations about vertical axes within a broad shear zone that includes the Tarera-Aiduna and Lowlands fault zones. West of the southern limit of the fold-and-thrust belt, the Tarera-Aiduna fault zone accommodates ~6 cm year^{-1} of left-lateral motion (McCaffrey & Abers 1991).

Volcanism

The New Britain arc is a relatively long-lived feature, with volcanism active during the Late Eocene, Late Oligocene, and Mio-Pliocene (Lindley 1988). It is thought to have initially formed as a result of southward subduction of the PAC plate under the AUS plate, with the Solomon Sea in the back arc (Yan & Kroenke 1993). Since 3.5 Mya, northward subduction of the Solomon Sea has resulted in arc magmatism in New Britain and back-arc basin formation in the Bismarck Sea.

Volcanism is active in the Woodlark Rift (Smith 1982) and Woodlark Basin (Taylor et al. 1995) (**Figure 3**). Presently dormant but well-preserved volcanic landforms are found on southeast Goodenough Island, southeast Fergusson Island, and Dobu Island. Geothermal fields are active at Deidei and Iamalele on Fergusson Island. Nearly all volcanic rocks in the Woodlark Rift possess some kind of subduction signature (e.g., light rare-earth element enrichment with negative Nb and Ta anomalies), even those demonstrably related to extension (e.g., Stolz et al. 1993) and even some Woodlark Basin basalts (Chadwick et al. 2009). The cause of volcanism west of the Woodlark Basin seafloor spreading center rift tip and on the Papuan peninsula is controversial. Active volcanism has been attributed to southward subduction of Solomon Sea lithosphere at the Trobriand Trough (e.g., Smith & Davies 1976, Smith 1982, Martinez & Taylor 1996) or rift-related decompression melting of subduction-modified mantle (Johnson et al. 1978).

On the Papuan peninsula, active volcanism historically occurs at Mt. Lamington, Mt. Victory, and Goropu, as well as in the Managalase Plateau volcanic field (Taylor 1958, Johnson et al. 1978, Ruxton 1999). The calc-alkaline stratovolcanoes Mt. Lamington, Mt. Victory, and Goropu show high Ba/La (>20) typical of arc volcanoes, and abundant hydrous phenocrysts suggest active subduction beneath the Papuan peninsula (Taylor 1958, Smith & Johnson 1981, Arculus et al. 1983). Yet Mt. Lamington lavas do not show expected ^{10}Be excess of recently subducted sediments (Gill et al. 1993). Goropu, a small vent on the Owen-Stanley fault, erupted in 1941 (Davies 1971). In the eastern central range of Papua New Guinea, Plio-Pleistocene volcanics were erupted through deformed Australian continental margin sediments (Cloos et al. 2005). In the highlands, some glaciated volcanoes were active in the Quaternary.

Tectonic Components

In discussing New Guinea tectonic components, we generally follow Crowhurst et al. (1996), Hill & Hall (2003), Cloos et al. (2005), and Davies (2012), focusing on major features common to most models. In this brief review, we simplify many aspects of, and controversies surrounding, the geologic evolution of New Guinea.

Tasman line. Along-strike variability in deformation styles is, in part, related to the location of the Tasman line (Scheibner 1974). This tectonic boundary separates Proterozoic craton (i.e., cold, thick, strong lithosphere) from regions to the east (i.e., younger, hotter, weaker lithosphere) that evolved since the Paleozoic in an active margin setting and/or were accreted to the continental margin. Although its exact location, especially in northwestern New Guinea, is poorly known (Hill & Hall 2003), it has important relevance for how deformation is accommodated as collisional orogenesis proceeds. In some regions, it is clear where Precambrian Australian craton has been reworked, such as in the Mapenduma anticline of the southwest Central Ranges of the New Guinea highlands block (Nash et al. 1993, Cloos et al. 2005). In other regions, Precambrian zircons provide clues regarding the extent of reworked Australian craton in New Guinea. Such zircons have been found in the Late Miocene Porgera intrusive complex in the New Guinea fold belt (Munroe & Williams 1996), in Permian metasediments in the Kubor and Bena Bena blocks of the central highlands (van Wyck & Williams 2002), and in Plio-Pleistocene sediments in the Trobriand Basin of the Woodlark Rift (Baldwin & Ireland 1995).

Passive margin to foreland basin. The southern lowlands—an aseismic, stable platform of low topography (<100 m) on the northern part of the AUS plate—are underlain by a ~1,000-km-wide zone of thinned continental crust formed during Triassic rifting (Veevers et al. 1991, Cloos et al. 2005). From Mesozoic–Paleocene, New Guinea occupied the leading northern edge of the AUS plate as a passive margin, during which a carbonate platform developed (e.g., Pigram et al. 1989, Hill & Hall 2003). The Eocene-to-Pliocene carbonate platform of southern New Guinea records platform evolution and its ultimate demise during the transition from passive margin to foreland basin. Collision of the northern edge of the AUS plate with several arcs, oceanic plateaus, and microcontinents (Pigram & Davies 1987, Davies 2012) led to the formation of the fold-and-thrust belt and the associated foreland basin to the south of the collision zone (Pigram et al. 1989, Quarles van Ufford & Cloos 2005).

Fold-and-thrust belts of the New Guinea highlands block. The New Guinea highlands block comprises a mountain belt that is ~1,300 km by 100–150 km. Reaching its highest elevation (4,884 m) at Puncak Jaya (also known as the Carstensz Pyramid), the western fold-and-thrust belt in the province of Papua in the Republic of Indonesia is generally higher in elevation (with peaks >3,000 m) than its Papua New Guinea counterpart. The western fold-and-thrust belt incorporates limestones as young as 15 Ma that were folded and uplifted together to form a mountain range since ~12 Mya (Cloos et al. 2005). Late Miocene collisional orogenesis involved reactivation of early Mesozoic(?) normal faults to form the Mapenduma anticline, a fold-and-thrust structure that exposes Australian basement and Paleozoic-Cenozoic strata. Shortening in the western fold-and-thrust belt is thought to have ceased by ~4 Mya, transitioning into minor northwest-striking sinistral strike-slip faulting that concentrated along the Yapen fault at ~2 Mya, but convergence is still ongoing in the eastern part of the Central Range (Cloos et al. 2005).

In contrast to the western fold-and-thrust belt, the eastern fold-and-thrust belt of Papua New Guinea is largely underlain by weaker, Paleozoic accretionary crust. This crust includes low-grade Permian to early Triassic metasediments intruded by Middle Triassic and Early Jurassic granites and overlying Jurassic to Tertiary sediments (Hill & Hall 2003). Following Late Oligocene–Miocene arc-continent collision (i.e., Melanesian arc collision), most folding and thrusting occurred in the Late Miocene–Holocene (Hill 1991, Hill & Hall 2003). Abers & McCaffrey (1988) estimated the rate of convergence in the Papuan fold-and-thrust belt to be 4–9 mm year^{-1}, assuming thickening occurred since 10 Mya.

Melanesian arc-continent collision is oblique, resulting in diachronous collisional orogenesis within the fold-and-thrust belts, beginning in the west and moving eastward (Cloos et al. 2005). The eastern fold-and-thrust belt is at the stage that the western part was at ~4 Mya. Other important considerations are that the New Guinea highlands block is accommodating AUS-PAC oblique plate convergence and collision by partitioning strain into (*a*) thrusts that build topography and internally deform by creating the fold-and-thrust belt to the south, and (*b*) sinistral strike-slip faults to the north within the Mobile Belt. From the Miocene to the Pliocene, deformation likely propagated southward (in-sequence thrusting), involving both thin-skinned deformation, reactivation of extensional faults, and generation of large basement-cored anticlines (Hill 1991, Hill & Hall 2003).

Ophiolites of New Guinea's mobile belt. Four major ophiolite belts (i.e., obducted and accreted oceanic lithosphere) (Davies & Jaques 1984) lie within the Mobile Belt (**Figure 2**). Although their ages, origin, and timing of obduction are debated (see Hill & Hall 2003), all of them occur sandwiched between fold-and-thrust belts to the south and accreted arcs to the north. The Papuan Ultramafic Belt ophiolite (PUB), the April ultramafics, and the Marum ophiolite belt represent oceanic lithosphere formed in forearc or back-arc basins and then emplaced during arc-continent collision above either north-dipping or south-dipping subduction zones (e.g., Hill & Hall 2003, Whattam 2009). Because these ophiolites structurally overlie high-pressure/low-temperature metamorphic rocks, as discussed below, northward subduction beneath island arcs, prior to obduction, is most likely.

The Late Cretaceous to Early Paleocene PUB is exposed on the northeast flank of the Owen-Stanley Range on the Papuan peninsula, where it is 16–21 km thick (Davies 1971, Davies & Jaques 1984). It extends >400 km into central New Guinea, with probable correlatives to the southeast in the Woodlark Rift (Davies & Smith 1971, Davies 1980b). Amphibole $^{40}Ar/^{39}Ar$ plateau ages from the metamorphic sole of the PUB are $\sim$58 Ma old, interpreted to result from cooling following obduction (Lus et al. 2004) due to collision with the northeast AUS plate at a latitude of 30°S. The PUB is continuous with oceanic lithosphere of the Solomon Sea, where it may be partly underlain by low-velocity material (Finlayson et al. 1977) inferred to represent subducted low-density crust (subducted AUS crust?). The extension of the PUB to the southeast remains controversial in part because only the Moresby Seamount is well dated. There, crustal extension is accommodated in part by normal faulting within latest Cretaceous to early Paleocene (66 Mya) oceanic crust (Monteleone et al. 2001). The presence of dolerite and gabbro in drill cores as far east as 152°E led to suggestions that the entire rift is overlain by $\sim$15 km of mafic and ultramafic rock (Martinez et al. 2001, Whattam 2009). However, mafic rocks recovered from drill cores are geochemically distinct from the PUB (Brooks & Tegner 2001), and seismic velocities (Ferris et al. 2006) do not show evidence for large thickness of mafic rocks in the rift.

The northeast-dipping dismembered Marum ophiolite is exposed along the northern flank of the Bismarck Range and to the south of the Ramu-Markham fault zone (Jaques 1981, Davies & Jaques 1984) (**Figure 2**). The Marum ophiolite is thought to be Early Jurassic and Paleocene in age (Page 1976, Hill & Raza 1999). Structurally beneath the Marum ophiolite are low-grade metamorphic rocks derived from continental slope flysch that transition into a Late Cretaceous–Eocene passive margin sequence of clastic sediments and limestones (Crowhurst et al. 1996).

The April ultramafic bodies occur in the foothills and northern flank of mountains within the Mobile Belt and south of the Sepik Basin in western Papua New Guinea (Dow et al. 1972, Davies & Hutchison 1982). Basalt and gabbro units have not been identified, but the ultramafic bodies were likely emplaced as thrust sheets, undergoing further dismemberment by east-west-striking sinistral faults (Davies & Jaques 1984). The April ultramafic bodies are thought to be Mesozoic or older and obducted in the Late Cretaceous (Jaques & Robinson 1977, Jaques et al. 1978).

The Irian ophiolite belt forms the northern flank of the New Guinea highlands (**Figure 2**). Composed of Jurassic oceanic crust and upper mantle, it represents uplifted and tilted forearc basement obducted prior to the Late Eocene (Weiland & Cloos 1996, Weiland 1999, Cloos et al. 2005).

Metamorphic rocks of the New Guinea mobile belt. As the leading edge of the northern AUS plate was subducted northward beneath the forearc of oceanic island arc(s), sediments and basalts were metamorphosed under high-pressure/temperature (high-P/T) conditions. Blueschists, eclogites, and lower-grade metamorphic rocks occur throughout New Guinea, primarily south of, and structurally beneath, ophiolites (**Figures 2** and **3**).

In the Woodlark Rift (**Figure 3**), prehnite–pumpellyite–to–greenschist and amphibolite facies rocks occur on the southern rifted margin, and blueschist–amphibolite–eclogite facies rocks are found in the lower plates of the D'Entrecasteaux Islands core complexes (e.g., Smith 1973; Davies & Warren 1988, 1992; Hill et al. 1992; Little et al. 2007). Late Miocene-to-Pliocene high-P/T metamorphism preceded diachronous exhumation from east to west, in the same direction as rift propagation (Baldwin et al. 1993, 2004; Monteleone et al. 2007). West of the active seafloor spreading center rift tip, the youngest (2–8 Ma old) (U)HP metamorphic rocks have been exhumed from mantle depths (>90 km) at plate tectonic rates (3–4 cm Ma^{-1}) (Baldwin et al. 2004, 2008).

On the Papuan peninsula, the Owen-Stanley fault zone separates the overthrust PUB from Kagi metasediments and Emo metabasites (e.g., Worthing & Crawford 1996, Davies & Williamson 2001) (**Figures 3** and **4*c***). The protoliths are largely Cretaceous sediments derived from the Australian margin. In general, metamorphic grade increases northward from prehnite–pumpellyite to blueschist (Worthing 1988), reaching the highest grade (i.e., amphibolite–granulite; Lus et al. 2004) at the base of the PUB. On the southeast Papuan peninsula, Emo metamorphic rocks are correlative with the blueschist-greenschist Goropu metabasalt (**Figure 4*b***), where they also occur structurally beneath the PUB (Davies 1980b). In the Suckling–Dayman Dome, pumpellyite-actinolite facies assemblages indicate exhumation from ∼25–35-km depths (Davies 1980b). These metamorphic rocks occur structurally beneath a greenschist facies mylonitic extensional shear zone (Daczko et al. 2011). In summary, remnants of a regionally extensive subduction complex are preserved in eastern New Guinea beneath the obducted PUB.

In the Sepik region of the north central New Guinea highlands, high-P/T metamorphic rocks [e.g., eclogite and glaucophane-epidote rocks (Ryburn 1980) and Alife blueschists (Weiland & Cloos 1996)] are associated with the April ultramafic rocks (**Figure 2*b***). Glaucophane K-Ar ages are 40 and 45 Ma, and phengite K-Ar ages are 25 and 28 Ma (Weiland 1999). These ages suggest Paleogene and possibly Early Miocene (Page 1976) subduction-zone metamorphism within a north-dipping subduction complex (Ryburn 1980) that extended into western Papua New Guinea. From north to south, metapelites and metabasites decrease in metamorphic grade (i.e., eclogite, lawsonite blueschist, prehnite-pumpellyite).

Further west, in the province of Papua in the Republic of Indonesia, blueschist and eclogite localities include Eocene metabasite blocks in the Yeleme quarries (Pétrequin & Pétrequin 1993, Weiland 1999) and greenschists and blueschists in the Efar-Sidoas Mountains (van der Wegen 1971). In the Cyclops Mountains, garnet-bearing greenschist yielded phengite K-Ar ages of 21 Ma (Pieters et al. 1983, Weiland 1999). To the south, in the Ruffaer metamorphic belt, blueschists and eclogites were exhumed from depths of 25–35 km (Weiland 1999). These high-P/T, as well as lower-grade, metamorphic rocks of the Ruffaer metamorphic belt may have been exhumed from a northeast-dipping subduction zone via normal faulting near the contact with the Irian ophiolite (Weiland 1999).

In the Bird's Head region, eclogite was exhumed during east-west extension in the Lengguru fold-and-thrust belt on the Wandamen peninsula (Bailly et al. 2009, de Sigoyer et al. 2010)

(**Figures 2*b*** and **4*e***). In the fold-and-thrust belt, two superimposed subduction complexes formed <11 Mya (Bailly et al. 2009).

Magmatism in New Guinea. As the AUS plate moved northward during the Cenozoic, its leading edge collided with island arc(s), and northern New Guinea has long been considered as the archetypal example of arc-continent collision. Although most models propose northward-dipping subduction of oceanic crust north of the AUS plate to form an arc, before slab rollback and collision of the northern AUS plate with that arc (see Hill & Hall 2003), the number and polarity of arcs accreted to northern New Guinea are controversial (see Quarles van Ufford & Cloos 2005 for a summary of six models). Accreted arcs now form numerous terranes within the Mobile Belt of New Guinea (Pigram & Davies 1987, Davies 2012) (**Figure 2**). In this section, we focus on aspects of two arcs included in most tectonic models of the New Guinea region—the Melanesian (or Bismarck) arc and the Maramuni arc. Enigmatic Late Miocene–Pliocene syncollisional magmatism is also discussed.

Melanesian arc collision. The mountain ranges that extend from the Huon peninsula toward the northwest (i.e., Sarawaged-Finisterre-Adelbert Ranges) consist of deformed Eocene pelites and Oligocene–Early Miocene volcanics, unconformably overlain by Miocene–Pliocene limestone (Jaques 1976). Most authors interpret these ranges as accreted island arc(s) (i.e., Melanesian arc of Crowhurst et al. 1996, Bismarck arc of Woodhead et al. 2010, Dabera-Bliri-Finisterre arc of Davies 2012) that formed above a northeast-dipping subduction zone from the Late Eocene to Early Miocene, with west-to-east oblique collision with the AUS plate margin beginning ~25 Mya and still ongoing (e.g., Jaques & Robinson 1977, Abbott et al. 1994, Hill & Hall 2003, Davies 2012). Some authors argue for a subduction polarity reversal following arc collision (e.g., Hamilton 1979); others, for south-dipping subduction prior to arc-continent collision (e.g., Cooper & Taylor 1987). Accreted arc terranes include, from west to east, the Tosem block, the Weyland terrane, the Gauttier terrane, the Cyclops Mountains, the Bewani-Torricelli Mountains, the Amanab block, the Finisterre Ranges, and the Huon peninsula (e.g., Pigram & Davies 1987, Hill & Hall 2003). Thrusting and sinistral strike-slip faulting have dismembered some parts of the arc terranes, translating them west (e.g., Cloos et al. 2005).

Maramuni magmatism. Miocene igneous rocks including high-K volcanics and granite, granodiorite, and diorite intrusives are broadly distributed to the south of the Owen-Stanley fault zone on the Papuan peninsula, in the Woodlark Rift, and in the northern and southern rifted margins of the Woodlark Basin (Smith 1972, 1973; Page 1976; Smith & Davies 1976; Ashley & Flood 1981). Further to the west, similar-age igneous rocks occur south of the accreted arcs, within the New Guinea Mobile Belt (Hill & Hall 2003), within allochthonous arc/forearc terrane in the Weyland overthrust (Dow et al. 1990), and at the westernmost end of the Irian ophiolite belt (Cloos et al. 2005). These igneous rocks have been interpreted to represent an arc formed by south-to-southwestward subduction of oceanic lithosphere beneath the rifted AUS margin from ~20–10 Mya, with the arc on, or near the leading edge of, the AUS plate (e.g., Davies et al. 1997). In this scenario, the Trobriand Trough in eastern Papua New Guinea may represent the relict subduction zone associated with the Maramuni arc.

Indeed, most Miocene intrusive and calc-alkaline volcanic rocks in the New Guinea Mobile Belt are attributed to the Maramuni arc (Page 1976), lying inboard (i.e., to the south) of the accreted Melanesian arc built on PAC plate oceanic lithosphere (discussed above). More broadly, Maramuni arc–related volcanism is considered to include all volcanism in the region, such as Miocene-Recent volcanism in the D'Entrecasteaux Islands, in islands such as the Amphlett Islands and Lusancay

Islands to the north, and in Mts. Lamington and Victory on the Papuan peninsula (Honza et al. 1987, Hegner & Smith 1992, Stolz et al. 1993, Taylor & Huchon 2002). On the southeast Papuan peninsula, high-K shoshonitic rocks of Miocene age intrude Eocene basalts (Smith 1972).

However, some authors do not consider Maramuni volcanism to be arc-related; instead, they associate it with partial melting following arc-continent collision and crustal thickening (e.g., Johnson et al. 1978, Johnson & Jaques 1980). Until processes responsible for generating Miocene magmatism are resolved (e.g., arc magmatism, delayed partial melting of subduction modified mantle, decompression melting of uplifting asthenosphere; Johnson et al. 1978, Woodhead et al. 2010), improvements in tectonic models will be limited.

Syncollisional magmatism. Late Miocene–Pliocene (i.e., 7.5–2.5 Mya) volcanic and correlative intrusive rocks, generally intermediate in composition, occur scattered over >1,000 km throughout the New Guinea highlands block, commonly near the highest elevations (e.g., Cloos et al. 2005). Although volumetrically minor, some of these magmatic rocks host world-class porphyry copper gold deposits, including the 6-Ma-old Porgera (Richards et al. 1990), 3-Ma-old Grasberg (MacDonald & Arnold 1994), and 1.5-Ma-old Ok Tedi (Rush & Seegers 1990) mining districts. Many of these magmatic rocks are mineralogically and chemically similar to arc rocks (McMahon 2001). The exception is the mafic alkalic Porgera intrusive complex, which has trace-element characteristics similar to those of basalts of intraplate settings, formed from partial melting of garnet amphibolite in the lower crust (Richards et al. 1990, Richards & Kerrich 2007). Cloos et al. (2005) proposed that Late Miocene–Recent syncollisional magmatism was due to asthenospheric upwelling and decompression of thinned mantle lithosphere between 7.5 and 3 Mya, as a result of delamination of the lithosphere.

To summarize, with the exception of the accreted Melanesian arc terranes, there appears to be a general pattern of magmatism younging toward the south from the Mobile Belt (20–10 Mya) to the fold-and-thrust belt (7.5–1 Mya) (e.g., Mason & Heaslip 1980, Hill & Hall 2003, Cloos et al. 2005). The origin and tectonic affinity of these magmatic rocks will remain controversial until their temporal and spatial evolution is more precisely known and linked to tomographic images of the mantle beneath New Guinea (see sidebar, Outstanding Questions).

The Bird's Head. The Bird's Head continental block is composed mainly of deformed Silurian-Devonian marine sediments intruded by Permian-Triassic granitoids (Pieters et al. 1983) and fault-bounded allochthonous blocks of Early Tertiary oceanic island arc affinity. The Bird's Head is actively deforming via anticlockwise rotation and subparallel sinistral strike-slip faulting (~8 cm year^{-1} relative to the AUS plate) (Pigram & Symonds 1991, Stevens et al. 2002, Bailly et al. 2009). Since ~6 Mya, the Bird's Head block has undergone 30–40° of anticlockwise rotation to form the Cendrawasih Rift (Charlton 2000).

Woodlark Rift and Basin. The Woodlark Rift (Hill & Hall 2003) (**Figure 2**) formed within heterogeneous lithosphere. Since the Late Miocene, a regionally extensive subduction complex has been exhumed, west of the active seafloor spreading center rift tip and on the southern-rifted margin of the Woodlark Basin (Pocklington Rise) (Baldwin et al. 2004, Monteleone et al. 2007, Baldwin et al. 2008) (**Figure 3**). In contrast, the northern-rifted margin (Woodlark Rise) comprises volcanic flows and pyroclastic material ranging in composition from basalt to rhyolite, with capping limestone (Trail 1967, Ashley & Flood 1981, Lindley 1994). The Pocklington and Woodlark Rises are conjugate rifted margins, separated by the Woodlark Basin seafloor spreading system since ~6 Mya (Taylor et al. 1995). Rifting reactivated structures within the plate boundary zone that previously separated subducted sediments and basalts from overthrust ophiolitic rocks of the PUB (Webb et al. 2008). Movement on extensional shear zones during the onset of rifting was associated with the exhumation of largely buoyant, previously subducted crustal material, from beneath mafic and ultramafic rocks, with some lower-plate rocks partially melting to form diapirs (Little et al. 2011).

Summary of major tectonic events. To summarize, the geology of New Guinea preserves evidence for arc-continent collision (Adelbert-Finisterre and other terranes), ophiolite obduction (PUB, Marum, April, and Irian), and subduction-zone metamorphism throughout the Mobile Belt. Ophiolites and high-P/T metamorphic rocks mark the suture between deformed AUS continental margin sediments in the New Guinea highlands fold-and-thrust belt to the south, and accreted volcanic arc(s) and microplates to the north (Davies 2012). The distribution of ophiolites and high-P/T metamorphic rocks (i.e., with an increase in metamorphic grade from south to north) within the Mobile Belt suggests that subducting slab(s) dipped to the north beneath oceanic island arc(s) prior to accretion. The tectonic affinity of Miocene-to-Recent magmatic rocks that generally young southward, intruding the Mobile Belt, then the fold-and-thrust belts, awaits further study.

LINKING DEEP EARTH PROCESSES TO LITHOSPHERIC AND SURFACE PROCESSES IN NEW GUINEA

The geology of the New Guinea region has been interpreted in a plate tectonic framework since the 1970s (e.g., Davies 1971, Curtis 1973, Hamilton 1979). The pronounced broad geoid high (and hence gravity high) over the southwest Pacific, and in particular the New Guinea region (**Figure 5**), is thought to result from high-density slabs in the upper mantle and related slab material in the deeper mantle (Hager 1984). However, how slab geometry and volume beneath New Guinea have evolved and impacted its tectonic evolution relies on further insights from tomography and models (e.g., Hall & Spakman 2003, DiCaprio et al. 2011) (see sidebar, Outstanding Questions). The potential role of hotspots and superplumes in the tectonic evolution of New Guinea is poorly understood. The southwest Pacific region, notably New Guinea, lies near the edge of the core-mantle boundary of the Pacific large low-shear-wave-velocity province (also known as the Jason LLSVP), a zone of plume generation (Burke 2011) (**Figure 5**). With the exception of a possible hotspot origin for the OJP and assisted opening of the Manus Basin, most New Guinea models ascribe ocean formation to back-arc extension driven by slab rollback, not taking into account potential plume involvement.

Effects of Cenozoic Northward Motion of the Leading Edge of the Australian Plate

The AUS plate underwent major changes in plate boundary forces as it progressively moved northward during the Cenozoic, with an increase in plate velocity since the Early Eocene (53–50 Mya) (Whittaker et al. 2007, Heine et al. 2010) (**Figure 1*b***). Cenozoic north-northeast motion of the AUS plate resulted in the Australian continent moving progressively into a region of higher geoid, thus contributing to an apparent north-side down tilting (Sandiford et al. 2009, Heine et al. 2010, DiCaprio et al. 2011). As the AUS plate moved northward, fragments of suprasubduction-zone (forearc) lithosphere (Metcalf & Shervais 2008), which formed in the earliest stages of magmatic arc formation, were subsequently accreted. These now comprise the ophiolites of the Mobile Belt. Exhumed blueschists and eclogites occur largely south of, and have been exhumed from beneath, obducted ophiolites. The world's youngest known (U)HP rocks (2–8 Ma old) in the Woodlark Rift, as well as eclogites in the Bird's Head on the Wandamen peninsula, were exhumed during

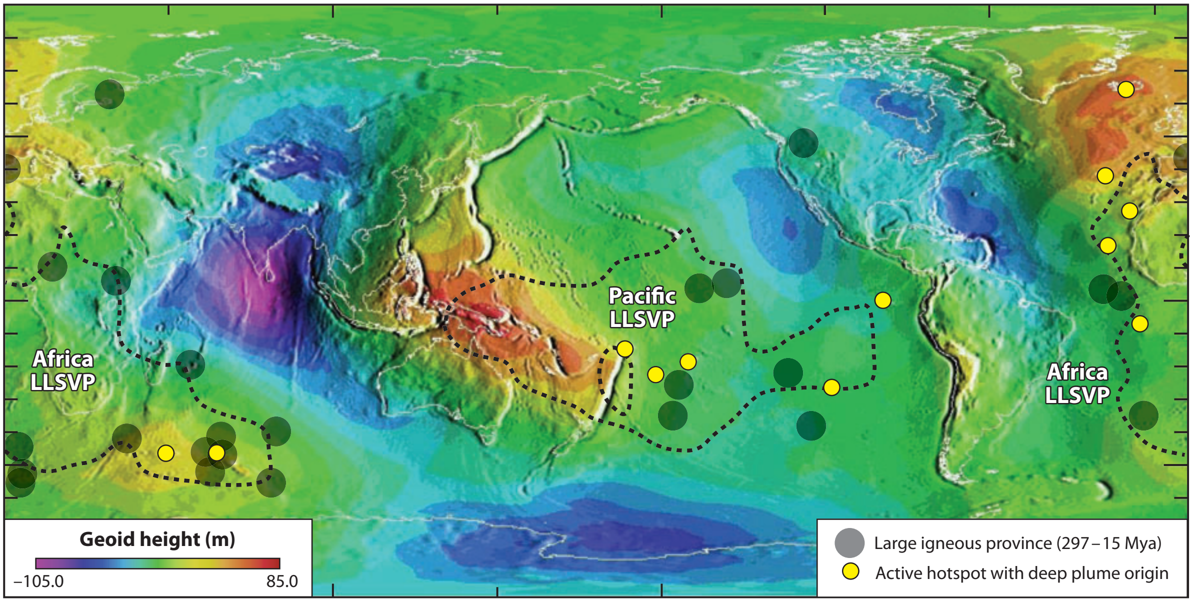

Figure 5

Global geoid map Mercator projection (with pseudoheight anomalies) generated from 15′ × 15′ geoid undulations from NIMA/GSFC WGS-84 EGM96 (figure modified from **http://earth-info.nga.mil/GandG/wgs84/gravitymod/**). Note the obvious geoid high over the New Guinea region. The dashed black lines represent the 1% slow contour of the ~2,800-km global composite tomography model derived from S-waves (SMEAN) (Becker & Boschi 2002), approximating plume generating zones and outlining the Pacific and Africa large low-shear-wave-velocity provinces (LLSVPs). Also shown are reconstructed (with respect to location at time of formation) large igneous provinces and active hotspots with a deep plume origin (modified from Torsvik et al. 2010).

a change from convergent to extensional tectonism involving microplate rotation within rapidly changing plate boundary zones.

Regardless of how many slabs there are and how their geometries evolved during the Cenozoic, it is likely that as the AUS plate moves northward, it progressively moved over subduction-modified mantle, potentially leading to partial melting of hydrated mantle and formation of igneous rocks with an arc geochemistry. The present-day along-strike differences in tectonic style in New Guinea, evident in schematic cross sections (**Figure 4**), are a function of variable along- and across-strike lithospheric composition, structure, heat flow, and rheology. It is striking that the extensional basins of northern Australia and southern New Guinea, floored by Proterozoic basement, have remained rigid within a long-lived convergent setting and managed to largely escape deformation during northward motion of the AUS plate.

Dynamic Topography in the New Guinea Region

The New Guinea highlands emerged from below sea level (i.e., foreland basin carbonates) and began to build topography beginning ~12 Mya (Cloos et al. 2005). Initially, deformation was accommodated via thin-skinned folding, but at ~8 Mya, thick-skinned basement-cored deformation continued to build an elongate orogen with elevations up to ~2 km. Cloos et al. (2005) propose that by 6 Mya, collisional delamination began, causing asthenospheric upwelling and magma generation via adiabatic decompression. Removal of lithospheric mantle resulted in isostatic uplift of the orogen to ~4 km in elevation. At ~4 Mya, denudation continued as strike-slip faulting became increasingly active. This dramatic buildup of topography in the New Guinea highlands (Weiland & Cloos 1996) as well as in the Finisterre Ranges (e.g., Abbott et al. 1994), in conjunction with the Pliocene emergent Indonesian maritime province, has influenced regional ocean circulation patterns and local climate (Cane & Molnar 2001, Karas et al. 2011). How much of this dynamic topography was driven by vertical coupling between the mantle and surface (Braun 2010) is presently unknown. However, in the Woodlark Rift, Abers et al. (2002) showed that the highest topography (in the D'Entrecasteaux Islands) is supported by buoyant mantle, and stream profile analysis of the metamorphic core complexes in the rift (Miller et al. 2012) indicate active transient rock and surface uplift above thin crust.

To summarize, virtually all plate tectonic and mantle processes are active in the rapidly evolving New Guinea region. Understanding the link between deep earth and surface processes in New Guinea provides a present-day view of processes that may have operated in the geologic past in other collisional orogens, such as the Oligocene western Alps (Malusà et al. 2011).

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