1. Introduction

The Papuan Peninsula-Woodlark Basin region (Figure 1) has been the focus of intense geological study given the important association of globally high extensional strain rates (rifting 10–20 mm/yr), seismogenic shallow-dipping (∼30°), and low-angle (22°) normal faults, the rapid emplacement (>10–20 mm/yr vertically) of the youngest (7–5 Ma) high- and ultra-high-pressure rocks in metamorphic core complexes and gneissic domes, the active breakup of a continent and the transition to seafloor spreading (Abers, 1991, 2001; Abers et al., 1997, 2016; Baldwin et al., 2008; DesOrmeau et al., 2014; Goodliffe & Taylor, 2007; Gordon et al., 2012; Little et al., 2011, 2019; Taylor & Huchon, 2002; Taylor et al., 1995, 1999; Tregoning et al., 1998; Wallace et al., 2004, 2014; Webber et al., 2018).

The close association and overlap of these processes in space and time, and their rapid evolution and legacy effects, has in several cases resulted in aspects of their current expression and inheritance/origin being interpreted differently. Some first-order questions include:

1. Is the Trobriand Trough the site of currently active subduction with a seismogenic slab and correlative Papuan arc volcanism? (Abers & Roecker, 1991; Cooper & Taylor, 1987a; Davies et al., 1984; Smith & Davies, 1976; Smith & Milsom, 1984)

2. Why are the geologically observed decadal strain rates in eastern Papua only about half of those inferred from seafloor spreading magnetic anomalies in the Woodlark Basin and, relatedly, how is the strain predicted ahead of the westward propagating seafloor spreading accommodated in the continental rift? Can the Papua-Trobriand Global Positioning System (GPS) data and Woodlark Basin spreading velocity data be better reconciled? (Abers et al., 2016; Kington & Goodliffe, 2008; Taylor, 1987; Taylor et al., 1999; Tregoning et al., 1998; Wallace et al., 2004, 2014)

3. Is the rapid emplacement of metamorphic core complexes and gneissic domes exhumed in the continental rift dominantly by horizontal (extension on low-angle normal faults) or vertical (diapir) tectonics, and related to former northward or recent/current southward subduction and magmatic arc processes? (Abers et al., 2016, 2002; Baldwin et al., 2004, 2008, 1993; Biemiller et al., 2019; Dacz-
Toward answering these and related questions, in this study, we revisit the geological, seismological, marine geophysical, and geodetic information to derive an improved plate kinematic description of the tectonics of the Papua-Woodlark region. We find that many of the disparate views can be reconciled by a better understanding of both the tectonic inheritance and recent changes in the configurations/velocities of plate boundary processes.

2. Previous Work

2.1. Geologic Background

In order to provide context for the neotectonics, we first briefly review the geological evolution of the Papuan Peninsula and its former eastward continuation in the now submerged rifted margins of the Woodlark and Pocklington rises (Figure 1; among others, Baldwin et al., 2012; Davies, 2012; Little et al., 2011, 2013, 2019, 2007; Martinez et al., 2001; Mizera et al., 2019; Webb et al., 2008; Webber et al., 2018)

Figure 1. Geology and tectonics of the Papuan Peninsula-Woodlark Basin region, as per Taylor and Huchon (2002) and references therein, including Davies and Jaques (1984) and Pinchin and Bembrick (1985). Yellow-filled circles locate active volcanoes that define the Papuan volcanic arc, including Mt. Lamington (L, which last erupted in 1951), Mt. Victory (V), and Moresby Strait (M) volcanoes. The Dawson Strait (D) volcanoes are comendites related to the Woodlark rift, which is on strike with the volcanic arc. Basic metamorphic core complexes include the Emo Metamorphics (E) and the Suckling-Dayman massif (S, D). The D’Entrecasteaux Islands (DE) include Goodenough (G), Fergusson (F) and Normanby (N). Other islands include Trobriand (T), Woodlark (W), Egum (E), and Misima (MI). MS, Moresby Seamount; PM, Port Moresby. Bolder lines emphasize our interpretation of the plate boundary faults. The red lines locate the neovolcanic axis of the Woodlark Basin spreading center; dashed lines show the limit of oceanic crust.
Contemporaneous T- or E-MORB terranes of SE Papua include the Milne Basic Complex tholeiites (Kutu Volcanics and Goropu Metabasalts) and Sadowa Gabbro (Osterle et al., 2020; Smith, 1982, 2013), Emo Metamorphics (Whattam, 2009; Worthing, 1988; Worthing & Crawford, 1996), and the gabbro-dolerite-basalt sequence that underlies Moresby Seamount and the Woodlark Rise to its north (Ashley & Flood, 1981; Brooks & Tegner, 2001; Monteleone et al., 2001; Taylor & Huchon, 2002). Seafloor spreading in the Coral Sea Basin (62–52 Ma; Bulois et al., 2017; Weissel & Watts, 1979) separated from NE Australia the Papuan and Louisiade microcontinental plateaus that then collided with the arc and mostly sutured the trench, thickly sedimented former remnants of which include the Aure, Pocklington, and Rennell troughs (Figure 2). Mineral ages as young as 24–22 Ma in the Owen Stanley Metamorphics suggest that the collision continued into and was completed by the early Miocene (Davies & Jaques, 1984; Davies & Williamson, 1998; van Ufford & Cloos, 2005).

Arc reversal and southward subduction at the Trobriand Trough produced Papuan (a.k.a. Maramuni or Trobriand) arc magmatism (granitic plutons and calc-alkaline volcanics) from the middle Miocene through the Holocene (Davies et al., 1984, 1987; Dow, 1977; Fitz & Mann, 2013a, 2013b; Hegner & Smith, 1992; Lackschewitz et al., 2001, 2003; Smith & Compton, 1982; Smith & Davies, 1976; Smith & Milsom, 1984; Stolz et al., 1993; van Ufford & Cloos, 2005). Whether the Trobriand Trough remains an active subduction zone has been questioned, based on the lack of seismicity evidence (Abers & Roecker, 1991). Furthermore, as the 1951 Mount Lamington andesite lacks a cosmogenic Be signature, it has been suggested that the Trobriand (Papuan) volcanic arc may represent the eruption of mantle enriched by prior subduction (Gill et al., 1993; R. W. Johnson et al., 1978). The Trobriand forearc developed an up to 5–7 km thick sedimentary basin that filled to sea level in the late Miocene from paleo-water depths >500 m (Davies & Jaques, 1984; Davies & Williamson, 1998; van Ufford & Cloos, 2005).
2.2. Gravity

The Papua New Guinea-Solomon Islands region of the western Pacific displays a rapidly evolving mosaic of microplates between the major Pacific and Australia plates that are obliquely converging at ~110 mm/yr (Tregoning et al., 1998; Wallace et al., 2004). The present and former subduction trenches in the region are marked by large negative free air gravity anomalies (FAAs, Figure 2). These include the New Britain Trench, accommodating subduction of the Solomon Sea lithosphere to the north beneath New Britain and Bougainville, and the San Cristobal Trench, accommodating subduction of the Australia Plate beneath Guadalcanal and San Cristobal. In between, where the young Woodlark Basin is being subducted northeast beneath the New Georgia Group of the Solomon Islands, there is a deformation front but it lacks a flexed outer rise, bathymetric trench, and associated negative FAA (Martineau et al., 1999). The Trobriand Trough also has an associated negative FAA as well as a large positive (>200 mGal) FAA along the Trobriand outer forearc high, paralleled by a small negative FAA associated with the Trobriand forearc sedimentary basin (Figure 2; Davies et al., 1984; Fitz & Mann, 2013a, 2013b; Francis et al., 1987). Notably, there is no negative FAA associated with the northern Woodlark Rise. The Trobriand Trough, outer high, and forearc basin all terminate eastwards at the Nubara transform fault. The implication is that the Solomon Sea Plate is or has been subducted on three sides, but not on the fourth along the Nubara Fault, and hence that the initial opening of the eastern Woodlark Basin was not “back-arc” (Taylor et al., 2009; Weissel et al., 1982). The pronounced Trobriand outer forearc gravity high can be explained by this part of the forearc being out of isostatic equilibrium, resting directly on the Solomon Sea slab. Gravity anomalies that trend just south of east in the Solomon Sea (Figure 2) likely reflect its relict spreading center. The NW-trending negative FAA near 147°E, 13°S (Figure 2) marks the buried rift valley of the former spreading axis of the Coral Sea Basin (Weissel & Watts, 1979).
The thickly sedimented Aure Trough, Pocklington Trough and Rennell Trough, remnants of Paleogene northward subduction, are marked by negative FAA; likewise the North Solomon Trough, the site of southward subduction of the Pacific Plate prior to ca. 10 Ma (e.g., Cooper & Taylor, 1985, 1987b; Hamilton, 1979; van Ufford & Cloos, 2005). East-trending rift basins in the Woodlark and Pocklington rises are marked by less-positive FAAs, as is the stair-step axis of the spreading center in the eastern Woodlark Basin (Figure 2; Martinez et al., 1999). Other negative FAAs are associated with rift basins, such as in Goodenough Bay, as well as with sedimentary basins among the Solomon Islands (Figure 2). The northern Australia plate boundary is traced by river valleys and a relative gravity low down the axis of the Papuan Peninsula.

2.3. Seismicity and Arc Volcanism

Early studies of earthquake seismicity and focal mechanisms identified the primary plate boundaries of the Papua New Guinea-Solomon Islands region, albeit not without some varying interpretations (Abers & Roecker, 1991; Cooper & Taylor, 1985, 1987a; T. Johnson & Molnar, 1972; Pascal, 1979; Ripper, 1982; Weissel et al., 1982). In Figure 3, we show an updated view, using relocated hypocenters from the International Seismological Center (ISC)-EHB catalog (Engdahl et al., 1998, 2020; Weston et al., 2018), with Centroid Moment Tensor focal mechanisms for shallow (<40 km depth) earthquakes (Dziewonski et al., 1981; Ekström et al., 2012) plotted at ISC-EHB locations. Purple dots show microseismicity recorded during two PASSCAL experiments in SE Papua (Abers et al., 2016; Eilon et al., 2015; Ferris et al., 2006). Active volcanoes from the Smithsonian catalog (Global Volcanism Program & Venzke, 2013) are located with yellow triangles. Cross-sections of seismicity and topography along profiles A, B, C, and N are shown in Figure 4. Boxed area locates seismicity projected onto orthogonal profiles C and N in Figures 4c and 4d, respectively. Additional topographic profiles 1 and 2 are shown in Figure 8. Wadati-Benioff zone contours are shown at 50 km depth intervals for the western half of the Figure (and inset map in Figure 4).

Subduction at the New Britain and San Cristobal trenches is characterized by normal faulting on the incoming plate, shallow thrust faulting beneath the forearc, and steeply dipping Wadati-Benioff zones extending to depths exceeding 200 km, overlain by active arc volcanoes (Figures 3 and 4). No bathymetric trench, nor outer-rise flexure and normal faults, nor deep NE-dipping Wadati-Benioff zone, occur where the young crust and lithosphere of the Woodlark Basin are being subducted beneath the western Solomon Islands and their many arc volcanoes (Figure 3, also see Chen et al., 2011). Large shallow thrust earthquakes do occur,
Figure 4. Top four panels (a–d): cross-sections of seismicity and topography along profiles A, B, C, and D located in Figure 3 and inset map. Seismicity is projected from within a zone 50 km wide on either side of the profiles, except for profile A which is from within a zone 100 km wide on the NW side only and for profile D which is from within a zone 80 km wide on either side. Bottom two panels (e and f): cross-sections of seismicity and topography along profiles C and N, such that all the seismicity within the rectangular box delimited by C-C’ and N-N’ is projected onto the two orthogonal profiles. Inset map (g) locates the seismicity cross-sectional profiles and the Wadati-Benioff zone contours at 50 km depth intervals on the top of the subducted lithospheric slab. NBT, New Britain Trench; RMF, Ramu Markham Fault; TT, Trobriand Trough.
such as the $M_w$ 8.1 earthquake of April 1, 2007 (Furlong et al., 2009; F. Taylor et al., 2008) and the $M_w$ 7.1 earthquake of January 3, 2010 (Newman et al., 2011), both of which generated tsunamis.

There are also two clusters of east-trending strike-slip earthquakes east of Simbo Transform and south of the San Cristobal “Trench” that suggest active deformation, possibly by left-lateral bookshelf faulting reactivating ~E-W abyssal hill faults. It is likely that this easternmost Woodlark Basin is a picoplate or deformation zone such that the north-trending Simbo Transform reflects motion between the Woodlark Plate and this picoplate rather than the current NNW motion relative to the Australia Plate. This area has been rotated counterclockwise, not unlike the rotation of the Magdalena microplates associated with attempted ridge subduction off Baja California (Michaud et al., 2006).

Adjacent to the ridge subduction slab window a south-dipping remnant Pacific slab is seismically active to depths of 200 km and may source Savo arc volcano (Cooper & Taylor, 1985, 1987b). But the dominant arc volcanism is in the New Georgia Group of the Solomon Islands, including in the forearc (e.g., on Rendova and Kavachi) and even south of the trench on the incoming plate (on Simbo Island and Coleman and Kana Keoki seamounts) (Chadwick et al., 2009; Crook & Taylor, 1994; R. W. Johnson et al., 1987; Konig et al., 2007; Taylor & Exon, 1987). Seismicity beneath these volcanoes is shallow and diffuse—not unexpected given how readily thin young oceanic lithosphere may be resorbed into the mantle. It is likely that this arc volcanism owes its geochemistry and location to both ridge subduction and arc reversal.

3. Description of Current Plate Boundaries

3.1. Woodlark Basin Spreading Center

Marine geophysical studies over the past 4 decades have characterized seafloor spreading in the Woodlark Basin with increasing resolution—both on and off axis—such that it is one of the few ocean basins that is completely and systematically mapped with swath bathymetry, acoustic imagery, magnetic, and gravity data (Goodliffe & Taylor, 2007; Goodliffe et al., 1997, 1999; Martinez et al., 1999; Taylor, 1987; Taylor et al., 1995, 1999, 2009; Taylor & Huchon, 2002; Weissel et al., 1982). As the oceanic crust is also thinly sedimented, the seafloor fabric and magnetic anomaly age distribution are very well known. The basin opening from ~0.5 to 3.6 Ma west of Simbo transform, where both conjugates of the spreading are preserved and have not been rotated approaching the Solomon subduction zone, is well described by a single Euler pole at 147°E, 9.3°S, near Port Moresby (Taylor et al., 1999). A complexity in the basin evolution, however, is a recent (within the Brunhes Chron) change in spreading direction and rate, manifest in reoriented spreading segments and transforms as well as nontransform offsets (Figure 5, Goodliffe et al., 1997; Martinez et al., 1999; Taylor et al., 1995, 1999). Land GPS studies suggest that the present Australia-Woodlark (A-W) opening pole is somewhat further south and that the spreading rate may have slowed substantially (Tregoning et al., 1998; Wallace et al., 2004, 2014). The recognition that the Nubara Fault continues SW of Woodlark Island to a triple junction with the western spreading segments (Figures 1, 2, and 5) also means that the Trobriand Plate is separate from the Woodlark Plate (Davies et al., 1984; Taylor et al., 2009; Wallace et al., 2014). Furthermore, if the Trobriand Trough remains an active, albeit slow, site of subduction (see below), then the Trobriand Plate is also separate from the Solomon Sea Plate.

In this and the next three subsections, therefore, we describe the current plate boundaries in the region, together with their fault and fabric parameters that we have measured, in order to solve, with the GPS data, the best fitting multiplate kinematic description of the present tectonics.

The opening of the Woodlark Basin west of Simbo Transform has been organized in five spreading segments, with overlapping nontransform boundaries between segments 1 and 2, transform faults between segments 2–3 (Moresby) and 3–4 (Davies), and an eastward propagating spreading center that lengthened segment 4 at the expense of segment 5 (Figure 5; Goodliffe et al., 1997; Taylor et al., 1999). Following the reorientation of spreading direction during the Brunhes Chron, the Moresby and Davies transforms as well as the propagator developed short intratransform spreading segments (Figure 5). The axial rift valley in the eastern Woodlark Basin subdivided into several reoriented rift segments which are the locus of many normal and strike-slip earthquakes. In contrast, there is almost no teleseismicity nor focal mechanisms on the relatively inflated but slower spreading segments 1 and 2 in the western Woodlark Basin, which have overlapping spreading centers but lack transform faults (Figures 3 and 5). This difference in spreading
morphology is accompanied by the eastern basin having lower magnetization, thinner crust, and being regionally ∼500 m deeper than the western basin. Martinez et al. (1999) examined and modeled these characteristics as being the result of rift-induced secondary mantle convection in the western basin, with its thicker continental margins, but not in the east, giving the western basin characteristics elsewhere associated with faster (rather than slower, in this case) spreading.

Noting that the spreading fabric of segments 1 and 2 cannot be copolar with those of segments 3, 4, and 5, and given the observations above, we infer that spreading on segments 1 and 2 is somewhat oblique. We therefore use the abyssal hill and transform fault azimuths associated with segments 3, 4, and 5 as input to our inversion for the A-W opening pole (Figure 5, Table 1).

3.2. Nubara Transform Fault

A cluster of strike-slip earthquakes with aligned NE-trending nodal planes marks the Nubara Fault on the Woodlark Rise (Figure 6). The right-lateral focal mechanism for the largest of these, a magnitude 7 event in 1974, was first published in Weissel et al. (1982). Since that time, the seafloor bathymetry has been fully swath mapped, revealing three subparallel strike-slip fault strands (with small right-stepping releasing bends) on the central of which the earthquakes cluster (Figure 6, Taylor et al., 2009). The Nubara Fault continues to the SW past the southern end of the Trobriand Trough, across the Woodlark Rise to the southern side of the Egum Graben (Taylor & Huchon, 2002; Taylor et al., 2009) where there is a cluster of right-lateral strike slip and normal fault earthquakes (Figures 3, 5, and 6). We have measured the azimuths of the trace of the Nubara Fault at points along its length that are not releasing bends (Figure 6, Table 2). We fit small circles to these data to determine the location of the Euler pole between the Woodlark and Trobriand/Solomon Sea plates.

3.3. Trobriand Trough and Papuan Arc

Trench flexure accompanied by negative FAA gravity and normal faulting occurs on three sides of the Solomon Sea, at the Trobriand Trough as well as the New Britain Trench (Figures 2 and 3). A large shallow thrust earthquake (magnitude 7.3), under the Trobriand outer forearc at 150.1°E, 8.4°S generated a local tsunami that killed at least 30 people on March 6, 1895 (Figure 7; Letz et al., 2016). Although this historic
earthquake may not be as well located as Letz and coworkers surmise, there is no question that it occurred beneath the Trobriand forearc, given where the tsunami was and was not reported. The Trobriand Trough is thickly sedimented and has recently undergone convergence, as evidenced by large thrust sheets, spaced 5–7 km apart, forming the lower landward slope of the trough, ponding sediments behind them (Davies et al., 1987; Silver et al., 1991; Figure 7). Our bathymetry swath mapping on R/V Kilo Moana cruise 0418 traced the active deformation front of the Trobriand accretionary prism along its whole length west of 153°E (Figure 7). In addition to these features, subduction at the Trobriand Trough has produced the characteristic architecture of an arc-trench system, including an outer-arc structural and FAA gravity high, up to 5–7-km thick forearc sedimentary basin, volcanic front, and behind-the-front volcanoes, with arc volcanics dated from 15 Ma to present (e.g., Davies et al., 1984; Fitz & Mann, 2013a, 2013b; Francis et al., 1987; Pinchin & Bembrick, 1985; Smith & Milsom, 1984; Taylor, 1999).

The seismogenic zone under the Trobriand forearc is contiguous westwards with the southern limb of the doubly plunging seismogenic slab under the collision of the Huon Peninsula and Adelbert-Finisterre range with the New Guinea mobile belt of the north Australian margin (Figure 4; Hayes et al., 2018, their Figure 3; Davies, 2012). As Abers and Roecker (1991) pointed out, however, the seismicity deeper than 40 km and possibly illuminating a slab east of 148°E is quite patchy. We reproduce their stacked seismicity plots across and along strike (Figures 4d and 4e) with the benefit of 3 decades more teleseismic data, plus the microseismicity recorded during two PASSCAL experiments in SE Papua (Abers et al., 2016; Eilon et al., 2015; Ferris et al., 2006). Critically, for over ∼500 km, the various tele- and microseismicity patches south of the Trobriand Trough (Figure 4e) project in cross section onto one dipping seismogenic zone from crustal depths to ∼125 km (Figures 4d and 8). Furthermore, the depths to Moho under the Trobriand outer forearc, derived from the receiver functions of seismic stations on the low-relief islands, are so deep (42–49 km, Figure 8) that the deepest are most likely to come from the Moho of the subducted plate (Abers et al., 2002). The mantle P-wave velocity variations of Abers et al. (2002) are compatible with the slab evidenced by the seismicity, whereas the mantle S-wave velocity variations of Eilon et al. (2016), that incorporated 3-D seismic anisotropy, inexplicably do not image the subducted slab evidenced by the microseismicity that was located by the same experiments (Figure 8). The teleseismic and microseismic data together define a Wadati-Benioff zone associated with a south-dipping subducted slab, the top of which we have contoured at 50-km depth intervals, which is seismogenic to successively greater depths westwards (Figures 3 and 4).
Although we conclude that the Trobriand Trough and Papuan arc represents a Neogene subduction zone that remains active, the current subduction rates may be no more than a few millimeters per year (Kir-choff-Stein, 1992; Reed et al., 1988), which also may explain the lack of a cosmogenic Be$^{10}$ signature in the 1951 Mount Lamington andesite (Gill et al., 1993). As a means to bracket the possible rates, we develop plate kinematic solutions for two cases: with and without current subduction at the Trobriand Trough.

3.4. Papuan Peninsula and Woodlark Rift

For much of its length the Australia-Trobriand (A-T) plate boundary trends SE along the axis of the Papuan Peninsula as the double-stranded Owen Stanley Fault, separating the Owen Stanley Metamorphics (Kagi Metamorphics, Emo Metamorphics) from the Papuan Ultramafic Belt (Figures 1 and 2; Davies & Jaques, 1984; Davies & Smith, 1971; Pieters, 1978; Smith & Davies, 1976). East of $\approx$148.25°E, the plate boundary trends ESE and becomes increasingly extensional, including the Mai’iu Fault on the north side of the Suckling-Dayman massif, the Gwoira Fault to its east, and as the Goodenough Bay fault offshore the north coast of the Peninsula, eventually curving east and NE as a transfer fault crossing central Normanby Island (Figure 7; Dačzkco et al., 2011; Fitz & Mann, 2013b; Little et al., 2007, 2019; Taylor & Huchon, 2002).

**Figure 6.** Bathymetry, topography, and tectonics of the right-lateral Nubara transform fault and surrounds, showing focal mechanisms of earthquakes from the Centroid Moment Tensor catalog (Dziewonski et al., 1981; Ekström et al., 2012) plotted at their International Seismological Center-EHB locations—except for the magnitude 7 earthquake near 154°E, whose focal mechanism and location is from Weissel et al. (1982). Azimuths of Nubara Transform Fault segments that are not releasing bends bounding the Woodlark Plate are shown in dark blue, with dark blue tadpole fill for the Trobriand segment and white tadpole fill for the Solomon Sea segment (Table 2). Plate triple junctions are circled in aqua. EG, Egum Graben; MS, Moresby Seamount; TT, Trobriand Trough. Enlarged images of the bathymetry in the two boxed areas are shown in the middle and bottom panels, revealing the detailed trace of the right-lateral Nubara Transform Fault and its releasing bends.
Although geomorphology and other data indicate that these faults localize the principal plate boundary motion along the Peninsula (e.g., Webber et al., 2018), there are subparallel normal faults and microseismicity further south, including those that border the Milne Bay graben (Figures 7–9; Jongsma, 1972).

Marine geophysical and earthquake data indicate that there is another set of normal faults overlapping to the north, that extend west from the Australia-Trobriand-Woodlark (A-T-W) triple junction to the D’Entrecasteaux Islands, including from east to west, Normanby, Fergusson and Goodenough islands, beneath which the crust is regionally thin (Figure 9). These normal faults include grabens to the north and south of Moresby Seamount, and three subparallel asymmetric graben north of Normanby Island that extend west to Fergusson Island (Figures 1, 7, and 9; Goodliffe & Taylor, 2007; Taylor & Huchon, 2002), in what Abers (2001) and others subsequently have collectively termed the Woodlark Rift. A cluster of microseismicity continues WNW along the north coast of Fergusson Island (Figures 3 and 9; Abers et al., 2016; Eilon et al., 2015). The three gneissic domes of the D’Entrecasteaux Islands are sinistrally offset along trend by NE-striking transfer faults (e.g., Davies & Jaques, 1984; Figures 1, 7, and 9). Another microseismicity cluster marks a NNE-striking transfer fault off the east coast of Normanby Island, and yet another trends WSW from Goodenough Island to the Cape Vogel Peninsula (Figures 3 and 9; Abers et al., 2016; Eilon et al., 2015). Together, these observations indicate that there are overlapping zones of extension (north and south) and transfer faults (east and west) that bound tectonic blocks that comprise the D’Entrecasteaux Islands and Goodenough Basin (Wallace et al., 2014). There is a tongue of thicker crust that trends from beneath the Cape Vogel Peninsula ESE under the Goodenough Basin (Figure 9), which is itself being internally deformed, as evidenced by active normal faults imaged on seismic reflection data there (Fitz & Mann, 2013a, 2013b; Mutter et al., 1996). The thicker crust is evidenced by a greater Moho depth (measured by receiver functions, Abers et al., 2016) that is structurally below the hanging wall rollover of the listric Goodenough Basin Fault and thus not caused by it. Although it has been associated (Fitz & Mann, 2013a, 2013b) with lower crustal flow accompanying roll-back of the Trobriand slab, as proposed by Kington and Goodliffe (2008), that is an ad hoc rather than emergent characteristic of their model.

The easternmost A-T plate boundary segment is also the westernmost Woodlark Basin spreading segment 1a (Goodliffe et al., 1997; Taylor et al., 1999). The A-T-W triple junction, which we locate at 151.83°E, 9.78°S, is formed between spreading segment 1a, the Nubara Fault (along the southern edge of the Egum Graben), with the third arm being a nontransform offset between overlapping Woodlark spreading segments 1a and 1b (see Figure 2 of Goodliffe & Taylor, 2007, for details). Note that there is also a cluster of mostly strike-slip but also normal fault earthquakes on the southern margin, ahead of the dueling/propagating spreading

| Lon (°) | Lat (°) | Azimuth (°) | ± (°) | Source | Prediction (°) Case 1 | Case 2 | Difference Case 1 | Case 2 | Rate (km/Myr) Case 1 | Case 2 |
|--------|--------|------------|------|--------|-----------------------|--------|-------------------|--------|----------------------|--------|
| 152.028 | −9.733 | 69.3       | 1    | Transform | 65.07 | 67.39 | 4.23 | 1.91 | 9.72 | 9.73 |
| 152.247 | −9.641 | 65.44      | 1    | Transform | 63.57 | 65.86 | 1.87 | −0.42 | 9.73 | 9.75 |
| 152.411 | −9.536 | 57         | 1    | Transform | 62.33 | 64.59 | −5.33 | −7.59 | 9.71 | 9.73 |
| 153.087 | −9.274 | 62.67      | 1    | Transform | 57.83 | 59.98 | 4.84 | 2.69 | 9.83 | 9.84 |
| 153.376 | −9.114 | 57.6       | 1    | Transform | 55.76 | 57.87 | 1.84 | −0.27 | 9.86 | 9.89 |
| 153.561 | −9.004 | 57.6       | 1    | Transform | 54.42 | 55.35 | 3.18 | 2.25 | 9.88 | 4.99 |
| 153.892 | −8.824 | 53.4       | 1    | Transform | 52.10 | 52.79 | 1.30 | 0.61 | 9.93 | 5.02 |
| 154.203 | −8.531 | 48.61      | 1    | Transform | 49.44 | 49.86 | −0.83 | −1.25 | 9.90 | 5.00 |
| 154.466 | −8.251 | 42.24      | 1    | Transform | 47.04 | 47.22 | −4.80 | −4.98 | 9.87 | 4.97 |
| 154.823 | −7.866 | 43.18      | 1    | Transform | 43.74 | 43.59 | −0.56 | −0.41 | 9.85 | 4.96 |

Table 2
Input Azimuthal Data From Points Along the Nubara Fault and the Estimated Standard Error

Note. The last three columns present Case 1 and Case 2 azimuthal predictions, the difference from observations, and rate predictions at each point.
segment 2 (Figures 3 and 7). A campaign GPS station on Strathord Island in that region is moving ∼5 mm/yr to the NNE relative to Australia (STRA, Figure 2 of Wallace et al., 2014) also suggesting that not all extension is focused on the spreading segments but is partly distributed in the margins.

4. Plate Kinematic Models

Given the Australia, Woodlark and Trobriand plate boundaries described above, we seek the best fitting multiplate kinematic description of the present tectonics. Among other things, our goal is to determine whether the Papua-Trobriand GPS data and Woodlark Basin opening history can be better reconciled and to analyze the kinematics to improve our understanding of the tectonic processes. Given that the current subduction rates at the Trobriand Trough may be no more than a few millimeters per year (Kirchoff-Stein, 1992;
above the subducted slab.

the northern edge of the low velocity anomaly within the mantle wedge

DʻEntrecasteaux gneissic domes (G, Goodenough) are aligned and overlie

GB (Goodenough Bay). The frontal arc volcanoes (V, Mt. Victory) and

In the top panel, normal faults are labeled M (Maiʻiu), Gw (Gwoira), and

slab evidenced by the microseismicity located by the same experiments.

velocity variations (Eilon et al., 2015) but that do not image the subducted

is overlain on a projected cross-section from 150.4°E of mantle S-wave

from the Trobriand Tough (TT). In the bottom panel, the same seismicity

reproduces Figure 4d showing all the seismicity projected onto profile C.

and C located in Figure 3. The seismicity plot (middle and bottom panels)

represents Figure 4d showing all the seismicity projected onto profile C.

In the middle panel, the seismicity is overlain on a projected cross-section

from 150.8°E of mantle P-wave velocity variations (Abers et al., 2002).

The A-T plate pair has the most extensive data set (Table 3). We use the

campaign GPS results of Wallace et al. (2014), augmented by two fault

slip directions and the continuous GPS station at Port Moresby (Willis et

al., 2016; MORE, operational 01/93-03/02, 04/02-01/10, and 03/13-

06/13) to solve for the relative motion of the Trobriand and Australia

plates in the ITRF2008 reference frame (Figure 7, Table 3). One slip di-

rection we use is that of the mylonitic lineations on the north flank of

the Dayman dome (Daczko et al., 2011), the other is of the megascopic

grooves on the lower north flank of the Moresby Seamount shallow-ang-LE normal fault (Speckbacher et al., 2011). Unlike Wallace et al. (2014),

we do not attempt to solve for the detailed block motions and coupling

coefficients within the DʻEntrecasteaux Islands and Goodenough Basin

region of distributed deformation. Nor do we derive an Australia Plate

motion separate from the global plate solution that uses all the reference

sites on the Australia Plate (23 in Altamimi et al., 2011, vs. 11 in Wallace

et al., 2014). The difference between the two Australia Plate motions in

SE Papua amounts to Ve = 0.62 mm/yr and Vn = −1.21 mm/yr. Nor do

we distinguish from Australia a separate Papuan Peninsula block that

may represent minor (1–2 mm/yr) motion south of the designated A-T

plate boundary. Rather, we partition the GPS data into those along the SE

Peninsula within error of Australia Plate motion, those that consistently

define a separate Trobriand Plate to the north, and a third group from the

overlapping distributed extension zone in between that we do not invert.

For example, GPS site velocities on the south (MORA) as well as north

(WAIB and WATL) sides of Goodenough Island are well fit by the same

Trobriand Plate motion, which indicates that there is little/no relative ex-

tension between them (Table 3, Figure 7). Hence, the westwards transfer

of extension from the north side of the DʻEntrecasteaux Islands to the

south side of Goodenough Bay appears complete by the Barrier Islands

transfer fault (Davies & Jaques, 1984) that trends SW between Fergusson

and Goodenough islands (Figure 7). On the other hand, the three GPS

sites closest to the Trobriand Trough on the outer forearc high (KAWA,

LOS2, and GUA1) display components of westward motion that differ

from predictions of the best fitting Trobriand Plate motion (by 2.4, 3.6,

and 7.9 mm/yr, respectively, Table 3, Figure 7). They may result from

greater coupling to the subducted Solomon Sea plate, and/or differential

motion within the Trobriand subduction earthquake cycle, and/or for

LOS2 and GUA1 incomplete removal of coseismic displacements from

Reed et al., 1988), as a means to bracket the possible rates we develop

plate kinematic solutions for two cases: with the Trobriand Trough being

(1) inactive and (2) active—in which latter case, the Trobriand Plate is

separate from the Solomon Sea Plate.

Our kinematic models are constrained by the following data: GPS rel-

ative velocities, transform fault azimuths, ridge axis azimuths, and two

fault slip directions. Data utilized in our inversion are listed in Tables 1–3.

Table 1 presents data utilized for the A-W plate boundary. These data in-

clude both transform fault and ridge axis azimuths from spreading seg-

ments 3, 4, and 5. We exclude azimuths from segments 1 and 2 due to

their inferred spreading obliquity and overlap with possible continued

extension of the rifted continental margins. Data utilized for the Wood-
lark-Trobriand (W-T) plate boundary are listed in Table 2 and consist only

of fault azimuths along the Nubara Fault, exempting the releasing bends.

The A-T plate pair has the most extensive data set (Table 3). We use the

campaign GPS results of Wallace et al. (2014), augmented by two fault

slip directions and the continuous GPS station at Port Moresby (Willis et

al., 2016; MORE, operational 01/93-03/02, 04/02-01/10, and 03/13-

06/13) to solve for the relative motion of the Trobriand and Australia

plates in the ITRF2008 reference frame (Figure 7, Table 3). One slip di-

rection we use is that of the mylonitic lineations on the north flank of

the Dayman dome (Daczko et al., 2011), the other is of the megascopic

grooves on the lower north flank of the Moresby Seamount shallow-ang-LE normal fault (Speckbacher et al., 2011). Unlike Wallace et al. (2014),

we do not attempt to solve for the detailed block motions and coupling

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SE Papua amounts to Ve = 0.62 mm/yr and Vn = −1.21 mm/yr. Nor do

we distinguish from Australia a separate Papuan Peninsula block that

may represent minor (1–2 mm/yr) motion south of the designated A-T

plate boundary. Rather, we partition the GPS data into those along the SE

Peninsula within error of Australia Plate motion, those that consistently

define a separate Trobriand Plate to the north, and a third group from the

overlapping distributed extension zone in between that we do not invert.

For example, GPS site velocities on the south (MORA) as well as north

(WAIB and WATL) sides of Goodenough Island are well fit by the same

Trobriand Plate motion, which indicates that there is little/no relative ex-

tension between them (Table 3, Figure 7). Hence, the westwards transfer

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sites closest to the Trobriand Trough on the outer forearc high (KAWA,

LOS2, and GUA1) display components of westward motion that differ

from predictions of the best fitting Trobriand Plate motion (by 2.4, 3.6,

and 7.9 mm/yr, respectively, Table 3, Figure 7). They may result from

greater coupling to the subducted Solomon Sea plate, and/or differential

motion within the Trobriand subduction earthquake cycle, and/or for

LOS2 and GUA1 incomplete removal of coseismic displacements from
large Solomon Island earthquakes that occurred during the GPS campaign measurement period, the latter as discussed by Wallace et al. (2014).

5. Methodology

A grid search of the region (longitude-latitude) was performed to determine the statistically best fit pole location for each plate pair given the input data (Tables 1–3). We utilized a bootstrap method (Efron & Tibshirani, 1986) involving random resampling with replacement of data inputs and, for each of 5,000 iterations, calculated chi-square statistics (sum of squares of the difference between observation and prediction divided by the square of the standard deviation) for each grid point for each of three datasets separately. For the A-T GPS data, an additional grid search in omega (angular rotation) space was performed at 0.1°/Myr and 0.01°/Myr then refined to 0.001°/Myr to calculate predictions of the expected linear velocity (in eastings and northings) at each GPS station to locate the statistically best fit rotation rate for any longitude-latitude pair result. We then took each plate pair’s 5,000 results as a reasonable representation of possible pole error and calculated (95 percentile) contours of confidence using a density function and methodology described by Wilson (1993). Next, chi-square statistics for each grid point within the 95 percentile confidence regions were recalculated without resampling and summed across each possible triplet solution.
Given the spatial results for all three poles and the rotation rate results for the A-T pole, we derive the best fit rates for the A-W and W-T Euler poles using the Triple Junction Vector Rule and the Law of Sines. Given the directions of three vectors and magnitude of one vector at the A-T-W triple junction, we can derive the statistically best fit magnitude of the other two vectors. Then, we tested the fit of the resultant Euler pole triplets with (1) the data by summing the individual plate pair chi-square statistics and (2) the three-plate closure criterion:

$$A_{oT} = A_{oW} + W_oT$$
where $A_\omega T$ represents the vector describing motion between the Australia and Trobriand plates, $A_\omega W$ the Australia and Woodlark plates, and $W_\omega T$ the Woodlark and Trobriand plates, each represented by a rotation pole (longitude, latitude, and angular rate).

6. Results

Euler pole locations are plotted on Figure 10 in longitude-latitude space and angular rotation rate-latitude space with their associated 95 percentile confidence regions (Table 4). Predictions, differences from observations and estimated rates at input data locations for the three plate pairs are listed in Tables 1–3. Rates of motion along the plate boundaries are presented in Figure 11. The A-T plate pair resulted in the same Euler pole for both Case 1 and Case 2, likewise for the A-W plate pair at least up to the 100th place in angular rate. Therefore, whether or not the Trobriand Trough is active does not change the relative plate motion vectors at the A-T-W triple junction (Figure 11 insert). However, the difference between 1.988°/Myr and 1.983°/Myr in the angular rate for the A-W Euler pole is associated with a shift of the W-T Euler pole by half a degree in longitude and a 10th in latitude between Case 1 and Case 2.

Figure 10. Plot of Euler poles and associated 95 percentile confidence regions resulting from this (BT20) and other studies. Red stars and confidence regions show our Case 1 results, blue our Case 2 results and gold Wallace et al. (2014) results. Other stars and error ellipses represent as labeled—TL99 (gray): Taylor et al. (1999) and T98 (green): Tregoning et al. (1998).
6.1. A-T Plate Boundary

The most robust kinematic data set and cornerstone to our model comes from the A-T plate pair, whose GPS data provide the only rate constraint in our inversions. Using GPS rates and two lineations, we calculate the best fit current A-T Euler pole and angular rotation rate (147.6°E, 9.7°S, 2.560°/Myr) and confidence regions (Figure 10). The error of the A-T Euler pole is tightly constrained and the results of Wallace et al. (2014) are within it (Figure 10). Model predictions and differences relative to the A-T GPS data are presented in the same format as the campaign measurements (eastings and northings, Table 3). The average differences are −0.27 and 0.60 mm/yr in easting and northing, respectively. Vectors along the plate boundary change from compressional to extensional from west to east given the location of the pole (Figure 11) and range from −10 to 20 mm/yr. The resultant A-T Euler pole is also close to the Taylor et al. (1999) A-W Euler pole for 0.52–3.6 Ma. Further reflection on this is presented in the Discussion.

6.2. A-W Plate Boundary

Bathymetry swath mapping data of the Nubara Fault crossing the Woodlark Rise (Figure 6, Taylor et al., 2009) and of the eastern Woodlark Basin volcanic axis (Figure 5) collected since Taylor et al. (1999) allowed us to slightly refine picks of the current axial seafloor fabric and transform fault azimuths for spreading segments 3, 4, and 5 as compared to previous studies. Our Case 1 (147.4°E, 12°S, 1.988°/Myr) and Case 2 (147.4°E, 12°S, 1.983°/Myr) results differ only slightly in angular rotation rate but the difference allows for greater change between cases for the W-T plate boundary (Figure 10, Table 4). With the short spreading history of the current Euler pole, there is limited distance between azimuthal observations resulting in a stretched confidence region that extends along the trend of the spreading center. Note that the Taylor et al. (1999) and Wallace et al. (2014) A-W Euler poles lie along the same trend since they utilized the same azimuth data originally published by Taylor et al. (1999), which also included the maximum horizontal compressive stress on Mount Victory as a constraint (Figure 10, Table 4). The modifications to the spreading center azimuths combined with the GPS constraints on the A-T Euler pole resulted in a slower angular rotation rate for the A-W pole than Taylor et al. (1999), similar to the rotation rate of Wallace et al. (2014) (Figure 10). Our favored result (Case 2) produces spreading rates along the Woodlark Basin spreading center that systematically increase from 19 mm/yr in the west to 35 mm/yr in the east (similar to the results of Wallace et al., 2014; Figure 11).

6.3. W-T Plate Boundary

6.3.1. Case 1: Three-Plate Solution

The recognition of a separate Trobriand Plate from the Woodlark Plate (F. Taylor et al., 2008; Wallace et al., 2014) accompanied the recognition of the Nubara Fault as a transform fault with short right-stepping releasing bends. This allowed us to fit a small circle to the fault to identify a possible Euler pole. In the case where there is no active subduction at the Trobriand Trough, the Nubara transform fault would be the W-T plate boundary along its whole length. Observations of the Nubara fault azimuth over that ~450 km length reasonably well constrain the azimuth to the Case 1 W-T Euler pole (148.3°E, 1.7°S, 0.573°/Myr) and confidence regions (Figure 10). For comparison, the gold star in Figure 10 is the W-T pole calculated from the A-T and A-W poles of Wallace et al. (2014; Table 4), which fits the azimuths along the Nubara Fault less well. The

| Euler poles       | Lon (°) | Lat (°) | Rot. rate |
|------------------|---------|---------|-----------|
|                  | Case 1  |         |           |
| AUS-WDLK         | 147.4   | −12.0   | 1.988     |
| AUS-TROB         | 147.6   | −9.7    | 2.560     |
| WDLK-TROB        | 148.3   | −1.7    | 0.573     |
|                  | Case 2  |         |           |
| AUS-WDLK         | 147.4   | −12.0   | 1.983     |
| AUS-TROB         | 147.6   | −9.7    | 2.560     |
| WDLK-TROB        | 148.3   | −1.8    | 0.579     |
| WDLK-SOL         | 148.8   | −2.2    | 0.311     |
|                  | Calculated |       |           |
| PAC-AUS (ITRF08) | 184.2   | −60.6   | 1.080     |
| SOL-TROB         | 147.8   | −1.4    | 0.271     |
| Wallace et al. (2014) |         |         |           |
| AUS-WDLK         | 146.99  | −11.48  | 1.85 ± 0.29 |
| AUS-TROB         | 147.85  | −9.54   | 2.69 ± 0.13 |
| WDLK-TROB        | 149.61  | −5.24   | 0.84      |
| Taylor et al. (1999) |         |         |           |
| AUS-WDLK/TROB (0.52–3.6 Ma) | 147   | −9.3    | 4.234     |
| AUS-WDLK         | 144     | −12     | 2.437     |
| Tregoning et al. (1998) |         |         |           |
| AUS-WDLK         | 145.2   | −10.8   | 1.86 ± 0.03 |

Notes: Plate pairs include Australia-Woodlark (A-W), Australia-Trobriand (A-T), Woodlark-Trobriand (W-T), Woodlark-Solomon Sea (W-SS), Pacific-Australia (P-A), and Solomon Sea-Trobriand (SS-T). A-W/T labels the 0.52–3.6 Ma Euler pole, during which time the Woodlark and Trobriand plates may have been one. See Figure 10 for the error estimates of our and published Euler poles.
The sum of squares of W-T residual azimuths (predicted vs. measured) is 112 for our Case 1 and 1252 for Wallace et al. (2014). Our resulting W-T Euler pole for Case 1 produces rates along the Nubara Fault that range from 9.8 to 10 mm/yr (Figure 11, in which the depicted Case 2 W-SS rates northeast of the intersection with the Trobriand Trough would double, from 5 to 10 mm/yr for Case 1).

6.3.2. Case 2: Four-Plate Solution

Given the evidence, summarized above, that the Trobriand Trough is accommodating active subduction, albeit slowly, the Nubara Fault is a divided plate boundary with the Trobriand and the Solomon Sea plates. This is our preferred Case 2 solution, for which the sum of squares of W-T residual azimuths (predicted vs. measured) is 101. The portion of the Nubara Fault to the east of the Trobriand-Woodlark-Solomon Sea triple junction (153.49°E, 9.04°S) represents the Woodlark-Solomon Sea (W-SS) plate boundary and the portion to the west the W-T plate boundary. This four-plate solution has the relative motion between the Solomon Sea, Trobriand, and Woodlark plates being all nearly in the same direction at the triple junction (Figure 11 insert). Although the W-T Euler pole locations are similarly or better constrained than the A-W pole (Figure 10), we lack independent rate measurements for the Solomon Sea Plate, so we arbitrarily partition the W-T rate of motion along the Nubara Fault equally between the other two vectors (W-SS and Solomon Sea-Trobriand [SS-T]). The result for the Case 2 W-T Euler pole (148.3°E, 1.8°S, 0.579°/Myr) is presented in Figure 10 by the northernmost blue stars and confidence regions. The light blue stars and confidence regions symbolize the W-SS Euler pole (148.8°E, 2.2°S, 0.311°/Myr). Rates along the Nubara Fault between the Trobriand and Woodlark plates are the same as Case 1 (9.8–10 mm/yr) and are about half that (4.9–5 mm/yr) between the Woodlark and Solomon Sea plates. The SS-T Euler pole (147.8°E, 1.4°S, 0.271°/Myr) is calculated from these results and predicts rates of subduction along the Trobriand Trough that range from ∼3 to 5 mm/yr (Figure 11). The SS-T vectors are increasingly oblique to the west along the Trough, such that their orthogonal convergence rates would be even less, consistent with the expectations of Reed et al. (1988) and Kirchoff-Stein (1992). The full rates could vary from zero up...
7. Partitioning Spreading During the Brunhes Chron

Older seafloor fabric within the Brunhes Chron represents a primarily left-stepping ridge system in the eastern part of the basin until counterclockwise rotation of the spreading direction produced a transtensional system that resulted in more and shorter ridge segments and offsets (Figure 5). West of Moresby Transform, the ridge segments are spreading obliquely and are right-stepping, so the transtensional change resulted in increased nontransform offsets. This obliquely spreading part of the basin (segments 1 and 2), without transform faults, is not amenable to the following calculations.

To quantitatively address the changes in spreading rate and direction within the Brunhes Chron, we first selected conjugate features on either side of the spreading ridges (segments 3, 4, and 5) that represent the youngest fabric formed about the older pole described by Taylor et al. (1999). These pseudo-isochrons are represented in green in Figure 12. We found that our calculated A-W pole location (147.4°E, 12°S) could be used to reconstruct the pseudo-isochrons back together; however, our calculated modern-day angular rotation rate (1.98°/Myr) was not sufficient within the time constraints. To determine the average age of our pseudo-isochrons, we calculated the total distances between conjugate points and the Brunhes/Matayama boundary (Figure 12, green tadpoles) and, using the older pole, derived a best fit age of 0.45 Ma. We also measured the width of reoriented seafloor fabric associated with the axial valley and, using the current pole, derived the oldest age of the current spreading rate to be about 0.225 Ma (Figure 12, yellow bars). We then calculated the remaining distance between our two known Euler poles (distance from end of yellow lines to 450 Ka pseudo-isochrons). From this distance and the time constraints, we calculated the necessary spreading rates and derived the angular rate of 4.234°/Myr to explain earlier spreading rates. This angular rate is also the angular rate of the Taylor et al. (1999), 0.52–3.6 Ma Euler pole. Therefore, we suggest that the A-W...
Euler pole changed location at 450 Ka, while maintaining the faster angular rate, and then slowed to the current angular rate at 225 Ka. This is the simplest three-fold partitioning of spreading during the Brunhes Chron. It uses an instantaneous change in Euler pole locations and a subsequent instantaneous change in Euler pole angular rates. More gradual and/or complex changes between 450 and 225 Ka are possible, summing to the same results.

8. Discussion

While our model for the neotectonics of the region provides a good explanation of the current day plate motions, there are several questions still to be addressed about the past tectonics. For instance, what is the age of the A-T-W triple junction? Also, what caused the spreading changes in the Woodlark Basin? We calculate that the most recent change in spreading direction of the Woodlark Basin spreading center occurred at 0.45 Ma and we know that spreading was consistent about one pole prior to that back to at least 3.6 Ma (Taylor et al., 1999). Therefore, we expect the boundary forces to also be consistent during that time. The question to be answered is: What initiated the late reorientation and slowing of the seafloor spreading? We consider four main driving forces that could result in the reorientation of the Woodlark Basin spreading center, some of which have previously been considered by other authors (i.e., Ott & Mann, 2015; Wallace et al., 2014; Weissel et al., 1982):

1. Slab pull from the New Britain (western) section of the New Britain Trench
2. Increased resistance to convergence west of the Woodlark Basin pole of opening
3. Perpendicularity of ridge subduction beneath the Solomon Islands
4. Propagation of the Nubara Fault to form the boundary between the Trobriand and Woodlark plates

All or any one of these possibilities are plausible given the known constraints. Pull from the subduction at the New Britain Trench has long been hypothesized to be the primary reason for the opening of the Woodlark Basin (Weissel et al., 1982) but the history of subduction there is not very well known other than that it has produced a seismogenic slab to depths exceeding 600 km and that there is an eastward propagating arc-continent collision (Cooper & Taylor, 1987a; T. Johnson & Molnar, 1972; Silver et al., 1991). A change in that system could contribute to the change in spreading direction of the Woodlark Basin spreading center at 0.45 Ma. This is similar to the system in the Red Sea-Gulf of Aden where the Tethys subduction under Zagros and Makran, coupled with the eastward propagating continental collision of Africa with Eurasia, has been modeled in 3-D lab experiments as the primary driving force for spreading, albeit that its localization was augmented by a zone of lithospheric weakness generated by the Afar plume (Bellahsen et al., 2003). In their model, it is the contrast between collisional locking of the western portion of the trench and the continued slab pull of the eastern active portion of the trench that causes the separation and rotation of the Arabian plate from Africa—which is quite analogous to the New Guinea collision—New Britain Trench situation opening the Woodlark Basin.

The opening of the Woodlark Basin and Rift has been matched across its pole of opening (near Port Moresby, Figure 1, TL99 Figure 10, Taylor et al., 1999), scissor-like, by compression of the NW Papuan Peninsula region (Figures 3 and 11). This has been expressed by the inversion of forearc normal faults there into thrusts (Figure 1, Pinchin & Bembrick, 1985) and by the formation of the rear-arc Aure-Moresby foreland fold and thrust belt (Bulois et al., 2017; Ott & Mann, 2015). Changes to the resistance to this convergence, due for example to finite shortening and crustal thickening, could be an additional cause for changing the kinematics on the opening side of the pole.

Alternatively, the reorientation of the Woodlark spreading axes may position them more perpendicular to subduction at the (eastern) New Britain-San Cristobal Trench, as occurred on the ridge system subducting beneath Baja California (Michaud et al., 2006). It is also consistent with the rotation of the seafloor fabric east of the Simbo Transform, and of the southern Simbo Transform itself, that dates to ~2 Ma (i.e., prior to magnetic anomaly 2; Figures 5 and 12).

Whereas one or all of the above may have caused the change in A-W Euler pole location at 0.45 Ma, establishing the Nubara Fault boundary between the Trobriand and Woodlark plates may have allowed the subduction driving forces to more readily reorient the A-W spreading segments at 0.225 Ma. Creating the
W-T plate boundary and the A-T-W triple junction also allowed the Trobriand Plate to resume its prior to 0.45 Ma northward motion relative to Australia (Figure 11 inset), perhaps driven by trench suction. This is reflected in the similarity between the locations of the Taylor et al. (1999) 0.52–3.6 Ma A-W pole and the current A-T pole (Figure 10). That the Trobriand Trough convergence and Nubara transform faulting are both directed orthogonal to the slab pull from beneath the western New Britain Trench may help explain how such slab pull is transmitted across both plate boundaries to cause the AW Euler pole and the NW-vectors of Woodlark-Australia opening (Figure 11). Unquestionably there is at least one active plate boundary (the Nubara Transform Fault), and we suggest two (the Trobriand Trough), between the New Britain Trench and the Woodlark spreading center and rifts (Figure 11, also see Biemiller et al., 2019, their Figure 5 and/or Figure 1 of Wallace et al., 2014). Either slab pull can be transmitted across the near-orthogonal motion on this/these boundaries without producing extension there, or New Britain slab pull is not a primary driver of the Woodlark Basin opening, and one or more of the three other possible causes discussed above dominates.

Our improved understanding of the Papua-Woodlark region neotectonics also sheds new light on its past tectonics. In particular, having demonstrated the presence of a patchily seismogenic subducted lithospheric slab under the Trobriand forearc, various correlative features can be more easily understood. Notably, the middle Miocene through Holocene Papuan calc-alkaline volcanic arc and granitic plutons (Davies et al., 1984, 1987; Dow, 1977; Fitz & Mann, 2013a, 2013b; Hegner & Smith, 1992; Lackschewitz et al., 2001, 2003; Osterle et al., 2020; Smith & Compston, 1982; Smith & Davies, 1976; Smith & Milsom, 1984; Stolz et al., 1993; van Ufford & Cloos, 2005) have a ready proximal source in the mantle wedge above the subducted slab. Furthermore, Australian continental materials subducted northward into the mantle in the Paleogene (Baldwin et al., 2004, 2008; Hill et al., 1995; Little et al., 2011; Zirakparvar et al., 2014) were remobilized ∼7–1 Ma in diapirs rising from >90 km depth (Baldwin et al., 2004, 2008; DesOrmeau et al., 2014; Montelione et al., 2007), accompanied by extensive partial melting, ductile deformation and leucocratic-granodioritic magmatism (Gordon et al., 2012; Hill et al., 1995), and been emplaced in the D’Entrecasteaux Islands as gneissic domes (Baldwin et al., 1993; Davies & Warren, 1988, 1992; Hill, 1994; Hill & Baldwin, 1993; Korchinski et al., 2014; Little et al., 2007, 2011) along the arc volcanic front in the Woodlark rift (Figures 1, 3, and 7–9). This causative correlation substantially has been drawn previously in the tectonic models of Martinez et al., 2001, Little et al., 2011, Ellis et al., 2011, and Fitz & Mann, 2013a, 2013b. In contrast, the Suckling-Dayman metamorphic core complex (MCC) is behind the volcanic front on the Papuan Peninsula and is in the footwall of the low-angle Mai’iu Fault (Figure 7). It has been well described and modeled as formed by an extensional detachment MCC (Biemiller et al., 2019; Daczko et al., 2009, 2011; Davies, 1980, Fitz & Mann, 2013a, 2013b, Little et al., 2019; Mizera et al., 2019; Webber et al., 2018). But even there, the fact that Mount Suckling (Goropi) has the highest elevation in Papua (3,676 m), that no higher “breakaway peak” that characterizes MCC models (Biemiller et al., 2019) exists, and that its summit includes calc-alkaline granite and high-K monzonite (4–2 Ma, Osterle et al., 2020; Smith & Davies, 1976) indicates that magmatic intrusion and uplift, and not just extensional detachment tectonics, likely contributed to its (vertical) evolution (Lister & Baldwin, 1993; Lindley, 2014).

9. Conclusions

1. Subduction at the Trobriand Trough is occurring, very slowly, based on the following evidence: trench flexure accompanied by negative FAA gravity anomaly and normal faulting, correlative Papuan arc volcanism, presence of a south-dipping patchily seismogenic slab to depths of 50–125 km beneath the Trobriand forearc contiguous westwards with the southern limb to 200 km depth of the doubly plunging seismogenic slab under the collision of the Huon Peninsula with the New Guinea mobile belt, a tsunami-generating historic earthquake in the outer forearc, large thrust sheets, spaced 5–7 km apart, forming the lower landward slope of the trough and ponding sediments behind them, and high-resolution bathymetric mapping of the active deformation front along the length of the trough (Figures 2–4, 7, and 8)

2. Our plate kinematic model estimates current spreading rates along the Woodlark Basin spreading center to range from 19 to 35 mm/yr, subduction at the Trobriand Trough to range from 3 to 5 mm/yr, right-lateral motion along the Nubara Fault at up to 10 mm/yr, and motion along the Owen-Stanley Fault and
Woodlark Rift to range from −10 mm/yr (compressional) to 3–21 mm/yr (extensional), all reported from west to east (consistent with previous studies; Figure 11)

3. The Papua-Trobriand GPS data provide the recent temporal (decadal) resolution for this plate kinematic model. Given the substantially faster seafloor spreading rates in the Woodlark Basin, measured from conjugate magnetic anomalies back to 3.6 Ma, we estimate a change in pole location at 450 Ka (within the Brunhes Chron) followed by, based on abyssal hill and fracture zone trends, a decrease in spreading rate at 225 Ka (Figure 12)

4. The emplacement of gneissic domes in the continental rift is along the volcanic front and, like the proximal arc magmatism, is dominantly by diapiric processes originating in the mantle wedge above the south-dipping lithospheric slab subducted at the Trobriand Trough (Figure 8) (that mantle having been preconditioned by continental materials introduced from the south during Paleogene subduction). In contrast, the Suckling-Dayman MCC formed in the footwall of the Ma’iu extensional detachment, aided, and abetted by rear-arc calc-alkaline and high-K magmatic intrusion and uplift

**Data Availability Statement**

The additional swath bathymetry data used in this study are available for download from the Rolling Deck to Repository (R2R; https://www.rvdata.us/search/cruise/KM0418) and Pangaea (https://doi.pangaea.de/10.1594/PANGAEA.900113?format=html#download). The microseismicity data are previously published in Ferris et al. (2006), and Eilon et al. (2015), and are available through the IRIS DMC.

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