How continents break up: Insights from Papua New Guinea

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Abstract. The Woodlark Basin in the western Pacific forms a continuous system of active continental rifting involving well-developed seafloor spreading. Thin sediment cover in the basin and a dominantly nonvolcanic rift phase permit basement fabric and structures to be imaged by swath mapping and seismic reflection data in the continental and oceanic parts of the basin. Magnetic isochrons indicate a single Euler pole of opening for most of the basin history and allow us to infer the opening kinematics along the rifted margins. In agreement with rigid plate tectonic models, continental rifting initiated geologically synchronously (at ≈6 Ma) along the length of the protocontinents within a deforming plate boundary zone. Strain localization and seafloor spreading, however, developed in a time transgressive fashion from east to west within this zone of deformation. Spreading centers formed within the rheologically weaker protocontinental margins surrounded by stronger oceanic lithosphere in the Solomon and Coral Seas. The transition to spreading occurred after a rather uniform degree of continental extension: 200±40 km. Both early and late stage rifting involved high- and low-angle normal faults. We identify distinct styles in the transition from rifting to spreading which we refer to as nucleation, propagation, and stalling. These breakup styles impart varyingly discordant to discordant relationships between the adjacent oceanic and continental rift structures. Continental transform margins which are or were juxtaposed against the ends of spreading centers show no evidence for thermal uplift or igneous underplating. The initial spreading segments achieved much of their length at nucleation (within rift basins separated along strike by accommodation zones), with subsequent lengthening by spreading propagation into rifting continental crust. This early propagation, and the subsequent development of transform faults between initially nontransform spreading segments, produced rift and spreading segmentation boundaries that are not simply correlated. The spreading centers nucleated approximately orthogonal to strike on the opening direction but, as the protocontinents were oblique to this direction, nucleation jumps occurred in order to maintain the new spreading centers within the protocontinents. Thus stepwise spreading nucleation in order to maintain within a rheologically weak zone, rather than rupturing of the lithosphere by stress concentration at the tip of a propagating ridge axis, is the dominant form of the rifting to spreading transition in the Woodlark Basin.

1. Introduction

The transition from continental extension to initial seafloor spreading is one of the least understood aspects of continental breakup. Continental deformation is distributed over large areas rather than focused on the narrow plate boundaries typical of oceanic lithosphere [Molnar, 1988; Haines, 1998]. The geologic and dynamic processes by which distributed continental deformation is progressively localized as an oceanic spreading center extends into a continent are widely debated [Bonatti, 1985; Rosendahl, 1987; Martinez and Cochran, 1988; Hayward and Ebinger, 1996]. Although the propagating rift hypothesis successfully explains one mechanism by which new seafloor spreading centers can replace others within rigid oceanic lithosphere [Hey, 1977; Hey et al., 1980], the model of rift propagation through rigid continental lithosphere has been disproved, and several alternate kinematic models of rift propagation into deforming continental lithosphere have been proposed [Vine, 1982; Courtillot, 1982; Martin, 1984; McKenzie, 1986].

Three-dimensional dynamic models of continental breakup have been constructed that correspond to some of the kinematic models [Bassi et al., 1993; Dunbar and Sawyer, 1996]. By varying the thickness and composition of the crust, the geotherm, the extension rate, and the strength and distribution of preexisting weak zones, these models can reproduce many of the typical margin types, with variable amounts of margin extension and volcanism. Nevertheless, the detailed processes involved in continental breakup and the birth of an ocean remain topics of continuing debate. Questions persist concerning the dominant mechanisms of extensional strain: whether by pure shear or simple shear and on low-angle and/or high-angle normal faults. Major low-angle detachments have been documented in continental rift systems such as the U.S. Basin and Range [Wernicke and Burchfiel, 1982], but their origin is controversial (original low-angle normal faults versus rollover of initially high-angle faults?), and few have been recognized in passive continental margins [Lister et al., 1991].

Many current studies of continental breakup are hampered by a lack of accurate kinematic constraints on the debated
Plate 1. A shaded topographic image of the Woodlark Basin region. Black dots locate shallow focus earthquakes west of 155°E. Inset shows location relative to New Guinea and Australia. Key tectonic features are marked and identified in the legend. Convergence at the subduction zone with open teeth is an order of magnitude slower than at the subduction zone with solid teeth. Major spreading segments are labeled 1 through 5, and the three sections of segment 1 are labeled a, b and c. Normally magnetized chrons are identified from Figure 1. Abbreviations are E, Egum Atoll; F, Fergusson; G, Goodenough; GB, Goodenough Basin; MB, Milne Bay; M, Misima; N, Normanby; MS, Moresby Seamount; MT, Moresby Transform; PP, Papuan Peninsula; R, Rossel; ST, Simbo Transform; SI, Solomon Islands; T, Tagula; TR, Trobriand; V, Mt. Victory; and W, Woodlark. G, F, and N are the principle islands of the D’Entrecasteaux Islands.
dynamic processes and/or by trying to reconstruct pre
Tertiary extensional events that occurred on conjugate
passive margins that are now buried by thick sediments,
thermally equilibrated, and separated by seafloor whose
spreading history is imprecisely known. Studies of regions of
intracontinental extension, such as the U.S. Basin and Range,
the East African Rifts and the Aegean, are also limited
because, although extension has occurred recently by
comparison to most passive margin examples, it has not pro-
cceeded to the point of seafloor spreading.

In contrast, the western Woodlark Basin-Papuan Peninsula
region (Plate 1) exemplifies a continuum of active exten-
sional processes, laterally varying from seafloor spreading to
continental rifting. Because of the youth of the breakup and
the generally thin sediment cover the pattern of faulting and
the transition from block-faulted continental crust to rugged,
diffracting, oceanic crust are readily apparent in swath
bathymetry and seismic reflection data [Hill et al., 1984;
Taylor, 1987; Taylor et al., 1995]. Seafloor spreading magneto-
amanomalies identified in the basin indicate that the
formerly contiguous, eastward extensions of the Papuan
Peninsula (the Woodlark and Rocklington Rises) were sepa-
rated as the Woodlark spreading center penetrated westward
during the last 6 Myr (Plate 1 and Figure 1) [Weissel et al.,
1982; Taylor and Exon, 1987; Taylor et al., 1995; Goodliffe,
1998]. Organized spreading ends at 151.7°E. Further west,
extension is accommodated by continental rifting, producing
full and half grabens such as Milne and Goodenough Bays,
respectively [Milsom and Smith, 1975; Mutter et al., 1996],
and metamorphic core complexes on the D’Entrecasteaux
Islands and the Papuan Peninsula [Davies, 1980; Hill and
Baldwin, 1993; Baldwin et al., 1993]. The extension is
accompanied by crustal tensional seismicity between 9° and
10°S as far west as 148°E [Weissel et al., 1982; Aber, 1991;
Aber et al., 1997] and by comenditic (transitional basalt-
peralkaline rhyolite) series volcanism in the D’Entrecasteaux
Islands [Smith, 1982; Hegner and Smith, 1992; Stolz et al.,
1993].

Carey [1958] postulated an extensional origin for the
Woodlark Basin, which he termed the Louisianiad "sphenochasm": a triangular gap of oceanic crust separating
two cratonic blocks with fault margins converging to a point,
and interpreted as having originated by the rotation of one of
the blocks with respect to the other. Since this prescient
description in the paradigm of continental drift, many have
sought to determine exactly how cratonic blocks are ruptured
to form the faulted margins bounding oceanic crust.

Using side-scan, magnetic, and seismic data recently
collected from this classic region (for data distribution and
processing details see Goodliffe et al. [1999] and Wessel and
Smith [1995]), we illustrate four types of continent/ocean
boundary (COB: nucleated, propagated, stalled, transform)
that typify different mechanisms by which oceanic accretion
is juxtaposed against continental rifts. We document major
low-angle detachments as well as high angle normal faults in
the rifted continental margins. Both modes of extension are
active at different times and places within the rifted continen-
t and either mode may focus breakup. We show that
continental breakup proceeds by successive phases of sifting
localization, spreading center nucleation, spreading center
propagation (sometimes stalling), and then a lateral jump of
the spreading axis to the next site of localized rifting.
Accommodation zones focus spreading nucleation to the
centers of late stage rift basins where normal faulting and
crustal thinning have been maximized. There is no simple
correlation between seafloor spreading segmentation and
Plate 2. (a) Bathymetry and (b) magnetization shaded with topographic relief illuminated from the north of the central Woodlark Basin reconstructed to 1.6 Ma. From 3.2 to 1.9 Ma, spreading segment 3 propagated westward until it stalled at magnetic anomaly 2 time against the asymmetrically rifted northern margin (see seismic section MW93-320 in Figure 3) and then "jumped" 8 km southward. Slightly after 2 Ma, spreading segment 2 nucleated farther west, in a symmetric rift graben (see seismic section MW93-310 in Figure 3) locally intruded by volcanics (labeled "v") that was ahead, and offset 70 km south, of the propagator. Accompanying the change in extensional faulting style from north dipping detachments bounding asymmetric half grabens to high-angle normal faults bounding a symmetric graben (Figure 3), there was a change in the rift segmentation and the location of accommodation zones (labeled "az"). From 1.9 to 1.5 Ma, spreading segments 2 and 3 overlapped with a ~60 km offset and 1:3 overlap-to-offset ratio. Moresby Transform formed at 1.5 Ma by cutting through the intervening continental lithosphere and truncating the tips of the two segments, most notably the eastern end of segment 2 (Plate 1).
margin rift segmentation, nor is there evidence for thermal uplift or igneous underplating of transform continental margins as spreading axes pass by.

2. Continent-Ocean Boundary

Our primary observations derive from the region west of 154°45'E, where we have obtained total coverage swath bathymetry and acoustic imagery, and 5-nautical-mile-spaced underway geophysics crossings of the Woodlark Basin and its continental margins (Plate 1 and Figures 1-4 [Taylor et al., 1995; Goodliffe et al., 1999]). Farther east, similar cover age exists for the axial regions but does not extend to the margins. Even there, however, published seismic reflection, bathymetry, magnetic, and gravity data from meridional tracks crossing the basin every 15-25 km [Taylor, 1987] are sufficient to interpolate the COB location and to infer the early basin evolution.
Figure 3. Single-channel seismic sections, located in Figure 4, across the margins of segment 2 showing (MW93-310) the symmetric high angle normal faults of the conjugate continental margins separated by oceanic crust formed during chron 2 (identified on magnetization profile above), with younger crust deleted, and (MW93-320) the asymmetric half grabens and a low-angle normal detachment in the continental margin farther north. Three insets show enlarged sections. Vertical seismic scales are in seconds of two-way travel time. The vertical exaggeration of the seafloor is ~5.

In locating the COB we do not have the benefit of seismic refraction data nor widespread basement samples. Nor are the free-air and Bouguer gravity anomalies associated with the COB of short enough wavelength to be particularly useful (Figures 2 and 4). However, given the generally thin sediment cover in the Woodlark Basin, the pattern of basement faulting can be readily discerned in swath bathymetry and seismic reflection data (Plates 1 and 2 and Figures 2–5). The lined up abyssal hill fabric of the oceanic crust is typically quite distinct from the block-faulted fabric of the continental margins, which often exhibit fault sets of multiple orientations. This is especially evident along COBs formed by spreading propagation (e.g., east of Moresby Transform, Figures 2 (bottom), 4, 5 and Plate 2). Even where oceanic and continental faults are subparallel across the COB (e.g., west of Moresby Transform, Figures 2 (top), 4, and 5 and Plate 2), seismic reflection profiles show that the fault blocks of the stretched continental crust typically step down to, and are tilted away from, the basin and have dammed thicker sediments than common on the oceanic crust (Figure 3).

Moreover, the character of magnetic anomalies typically changes across the COB, from being strongly lined up and parallel to the abyssal hills to being lower amplitude and more irregular on the continental side. There are some instances such as segment 2, however, where large amplitude curvilinear magnetic anomalies parallel segments of fault blocks and are even approximately axisymmetric on the conjugate margins (Figures 1, 2 (top), 3, and 5 and Plate 2). This phenomenon has been observed in other rifted margins and interpreted to reflect "transitional" crust (i.e., stretched crust with structurally controlled intrusions) (Martinez et al., 1995; Whitmarsh and Miles, 1995; Manighetti et al., 1997). Last, the conjugate margins can be accurately reconstructed (e.g., Figures 3 and 5 and Plate 2) because the history of seafloor spreading is tightly constrained by the magnetic lineations and seafloor fabric (Figure 1 and Plate 1). Such reconstructions (Goodliffe, 1998) provide an added constraint on the location of conjugate COBs.

The combination of swath bathymetry, magnetic, and seismic reflection data, west of 154°45' E, indicate a sharp transition (<5 km wide at the seafloor) from oceanic crust to stretched continental or transitional crust (Figures 2, 3, and 5 and Plate 2) [Taylor et al., 1995]. Farther east, we have dashed the COB in Figure 1 and Plate 1 where we have interpolated its location between the more widely spaced geophysical track data and COB picks published by Taylor [1987].

3. Seafloor Spreading History and Euler Poles of Opening

The magnetics data presented in Figure 1, and their interpretation superimposed on illuminated bathymetry in Plate 1, reveal the history of seafloor spreading in the Woodlark
Figure 4. Geophysical data in the vicinity of Moresby Transform and its conjugate margins: (left) acoustic imagery superimposed on bathymetric relief, (middle) free-air gravity anomalies, and (right) Bouguer gravity anomalies. Gravity anomalies are contoured every 10 mGal and labeled every 50 mGal.

The basin has been opening since at least chron 3A.1, approximately 6 Ma, at which time the westernmost spreading segment was located east of 157°E (Plate 1). The Woodlark Basin spreading center grew discontinuously, stepping across Simbo Transform (156.5°E) about 3.6 Ma and Moresby Transform (154.2°E) about 2 Ma, to reach 151.8°E by 0.7 Ma. The 500-km-long spreading axis reoriented synchronously about 80 ka, producing the present discordance of the spreading axes with the Brunhes Chron boundary (Figure 1 and Plate 1) [Goodliffe et al., 1997]. In the following sections we refer to the spreading segments by number according to their configuration prior to the 80 ka reorientation, though we also discuss the subsections (a, b, and c) of segment 1 (Plate 1) [Taylor et al., 1995; Goodliffe et al., 1997].

The magnetic isochrons and isochronous seafloor fabric generated by spreading from 0.52 to 3.6 Ma (early Brunhes through anomaly 2A) are well described by a single Eulerian pole at 9.3°(±0.2)°S, 147°(±1°-2°)E, 4.234°/Myr. An exception occurs east of Simbo Transform, where the seafloor fabric and magnetic isochrons are not fitted by this pole (Plate 1). We infer that the area east of Simbo Transform was rotated anticlockwise together with seafloor (and a spreading segment) subducted beneath the Solomon Islands.

The fracture zones record a change in the pole of opening at 0.52 Ma (i.e., preceding the 0.08 Ma spreading axis reorientation) to 12°S, 144°E, 2.437°/Myr. This pole is located within the error ellipse of the position of the current Australia-Woodlark Euler pole (10.8°S, 145.2°E, 1.86±0.03°/Myr) determined by Tregoning et al. [1998] from two to five repeat observations of six regional Global Positioning System (GPS) stations but predicts opening rates over the last 0.52 Myr about 30% faster than the geodetic rates. Our pole correctly predicts the orientation (060°) of the axis of maximum horizontal compressive stress revealed by an alignment of six satellite cones on historically active Mount Victory and the colinear extinct volcano (Mount Trafalgar) to its ENE (Plate 1) [Nakamura, 1977; Smith, 1981]. Both poles predict slow (~10-15 mm/yr) left-lateral strike-slip motion along the Papuan Peninsula west of 148°E. An active, NW striking,
Figure 5. Structure of the central Woodlark Basin reconstructed to 1.0 Ma. See Plate 2.

left-lateral strike-slip fault mapped there between 147° and 148°E (Gira Fault [Davies et al., 1984]) may accommodate this motion.

4. Spreading Center Nucleation

The type example of spreading center nucleation within the western Woodlark Basin is spreading segment 2, where the COB is concordant with chron 2n, which marks the oldest adjacent oceanic lithosphere (Plates 1 and 2 and Figures 1-5). Younger oceanic isochrons, as a result of repeated minor reorientations of the spreading center, are subparallel (Figure 1 and Plate 1). The conjugate margins between Misima and Rossel Islands in the south and from Woodlark Island to Moresby Transform in the north are cut by numerous normal faults that form curvilinear horsts and grabens (Plates 1 and 2 and Figure 5). The southern margin segment is bounded to the east of Rossel Island by a SE trending aulacogen. Both margins shallower than 2000 m water depth are characterized by multiple rift basins each about 20 km wide by 50-100 km long. These basins and their bounding faults interfere near 153.5°E in inferred accommodation zones. The rift basins on the northern margin are asymmetric half grabens, with dominantly north dipping normal faults and successively rotated sedimentary sequences above south tilted fault blocks (Figure 3, bottom). At least one of the rift-bounding master faults is a low-angle normal detachment.

The pattern of faulting and rift basin segmentation on the segment 2 conjugate margins is strikingly different in water depths deeper than 2000 m (Plate 2 and Figure 5). There, the major block-bounding normal faults are high-angle (≥45°), spaced 2-7 km apart, symmetrically dip basinward, and sequentially step the continental basement down to 3.5 km below sea level (Figure 3, top). The northern margin is segmented by an accommodation zone at 153°15'5'E separating 100-km-long and 80-km-long half grabens to the east and west, respectively. The bounding half grabens on the southern margin that truncate the SE trending aulacogen are continuous for 140 km from 153° to 154°15'E. When reconstructed to their breakup geometry, the segment 2 conjugate margins form a continental rift basin 140 km long (Figure 5). Oceanic abyssal hill fabric and chron 2n are subparallel to the inner normal faults and maintain a constant total separation from them along at least the 40-km middle segment of the basin. Along strike toward either end, the oceanic isochrons gradually intersect the COB. Several centers of rift volcanism can be identified in the bathymetry and magnetics of both margins (v in Plate 2), but they are not so extensive as to mantle the topography of the major normal faults.

In the absence of deep crustal structure and drilling information we interpret these observations of the segment 2 margins and breakup as follows. The early rifting was spatially and/or temporarily distributed. Extension occurred on dominantly north dipping faults, some of which were low angle. The asymmetry perhaps reflects preexisting structural grain inherited from the Palaeogene Papuan orogeny [Rogerson et al., 1987]. Late stage rifting focused extension and formed one 140-km-long basin in which normal faulting was axisymmetric and high angle. This rift basin was delimited from others along strike by accommodation zones. Volcanoes locally, but not extensively, extruded lavas within the rift. Upwelling asthenosphere fed magmas into the thinning continental lithosphere and nucleated a spreading center along the central segment of the rift axis. This spreading center then propagated rapidly (~500 mm/yr) along axis to both ends of the basin, producing overall concordance between the subparallel rift and seafloor fabrics.

Concordance between oceanic isochrons and the COB is also seen in spreading segment 1 between 151°50'E and 152°30'E and in the failed spreading segment northeast of there (Plate 1). Except for large curvilinear normal faults bounding the latter to the north, the conjugate normal faults in the immediately adjacent continental margins are more numerous, closer spaced, and smaller offset than for segment 2. Further from the COB, both margins of segment 1 are cut by oblique normal faults. Mutter et al. [1996] interpret these fault blocks in the southern margin as rotated about a vertical axis in a migrating extensional relay zone caught between offset centers of spreading and rifting.

Concordance between oceanic isochrons and the COB is consistent with the sparse data available that crosses the eastern half of the segment 3 margins and all the segment 4 margins (Figure 1 and Plate 1). We infer that these COBs, like the segment 1 COBs, also formed by the nucleation of spreading segments. The eastern half of segment 3 nucleated in a rift basin whose western continuation can be seen in the well-surveyed portion of the northern margin at 8°50'S (Plates 1 and 2 and Figures 4 and 5). The 8°50'S rift continues westward to within 20 km of the transform boundary that parallels the Solomon Sea oceanic lithosphere. A similar structural geometry occurs at the western end of the segment 5 northern COB. The much stronger rheology of the oceanic lithosphere in the Solomon Sea is the probable cause for the locus of focused rifting and eventual breakup stepping southward into the rheologically weaker Papuan continental lithosphere [cf. Vink et al., 1984; Steckler and ten Brink, 1986].
5. Spreading Center Propagation

The type example of spreading center propagation into the Woodlark Basin margins occurs adjacent to the western half of spreading segment 3 (154.2°-154.6°E) where the abyssal hill fabric intersects the COB at 40°-50° (Plates 1 and 2 and Figures 1, 2, 4, and 5). Westward younging intersections of oceanic isochrons 2A and 2 with the margins reveal that the COBs are not isochronous but formed by westward propagation at ~93 mm/yr of spreading segment 3 between approximately 3.2 Ma and 1.9 Ma. The duplication of anomaly 2, and probably parts of anomaly 2A, on the northern side of the basin indicates a seafloor spreading history with one or more southward ridge “jumps,” though we do not have sufficient data from the southern side of the basin to allow us to determine the mechanism by which these jumps occurred. The bathymetry of the oceanic crust deepens toward the COB, suggesting that the propagating tip was relatively magma starved.

The fault fabric of the northern margin is discordant with the COB of segment 3 (Plates 1 and 2 and Figures 4 and 5). Unlike segment 2, the transition from continental to oceanic lithosphere is not associated with successive downfaulting and thinning of COB concordant fault blocks. Instead, the margin structures are complex and three-dimensional, with two dominant fault trends: E-SE and NE. The E-SE normal faults are concordant with structures farther from the COB and reflect the N-S stretching. The NE trending faults are spatially proximal to the COB and probably result from distributed shear in the overlap zone from 3.2 to 1.5 Ma between rifting/spreading segments 2 and 3. There are no volcanic cones or high-amplitude magnetic anomalies in the propagator margin, implying that volcanism and probably magmatism did not play a significant role in their evolution.

We infer that spreading segment 3 nucleated in the east, then propagated westward into continental lithosphere whose locus of rifting was offset to the south (subsequently becoming spreading segment 2). In the absence of local stretching or margin magmatism to accommodate the strain, the propagator crosscut the existing E-SE trending horsts and grabens formed by previous continental extension. The resulting COB pseudofaults are discordant to the oceanic isochrons and the northern (outer) pseudofault is discordant to the margin structures.

At the western end of segment 2, between 152°57'E and 153°10'E, chron 2n and younger abyssal hill fabric intersect the COB at angles of 210°-50°, indicating that this part of the margin also formed by slow propagation (30-65 mm/yr). Thus spreading segment 2 first nucleated, then rapidly (~500 mm/yr) propagated, and then propagated slowly to the western end of the rift basin, where it finally stalled (see section 6). Unlike the western half of segment 3, however, the spreading center propagation in segment 2 was centered on the axis of continental extension. The result is that in segment 2 the structures on the conjugate margins are subparallel and concordant with the COBs.

Since the 80 ka reorientation, the western end of segment 2 is in the process of propagating into the southern continental margin north of Misima Island (Plate 1). There is a graben immediately ahead of, and produced by, the propagating spreading tip, but regionally, extension is focused on spreading segment 1 to the north. If spreading propagation continues, the result will be a discordant southern COB similar to the segment 3 west northern margin.

6. Stalled Segments and Transform Boundaries

Along the southern COB of segment 2 at 152°57'E, 10°30'S (northeast of Misima) and at the western end of segment 1b (151.8°E), oceanic isochrons, seafloor fabric, and continental margin structures are nearly perpendicular to the COB (Plate 1). In the former case, the COB is characterized by an abrupt increase in depth toward oceanic crust. In the latter case, the continental margin west of segment 1b forms a rift basin deeper than the adjacent oceanic crust, implying that the continental crust was substantially thinned. These boundaries represent stationary periods during which migration of the spreading system into the continental margin stalled. In the case of segment 1b the stationary phase recently ended with the nucleation of the westernmost segment (1a) in the last 0.1 Myr. Immediately west of segment 1a, extension is currently localized on a normal fault that dips NNE at 77°-43° from Moreshy Seamount, a continental block with greenschist facies metabasic basement (Figure 6) [Taylor et al., 1995, 1999; Mutter et al., 1996; Abers et al., 1997]. Thus, at present, continental breakup is focused on an asymmetric rift basin bounded by a low-angle normal detachment whereas just after 2 Ma, segment 2 breakup occurred along a symmetric rift basin bounded by high-angle normal faults.

The westward propagation of segment 3 stalled at about 1.9 Ma when it overlapped with segment 2 which had nucleated 70 km to the south (Plate 2). The large, overlapping, nontransform offset between spreading segments 2 and 3 persisted until 1.5 Ma, when Moreshy Transform joined the two by cutting through the intervening continental lithosphere and truncating their tips, particularly the eastern end of segment 2 (compare Plates 1 and 2) [Taylor et al., 1995]. The Moreshy Transform COB is approximately perpendicular to oceanic isochrons, seafloor fabric, and continental margin structures and has a sharp depth offset. Similar characteristics define the northward continuation of this north striking COB, which is a stalled COB segment (formed prior to the transform fault) where it is adjacent to the western end of the double magnetic anomaly 2 (Plate 2).

Interestingly, it is west of this stalled COB segment that some of the largest offset faults, including at least one low-angle detachment, occur in the northern margin (Figures 3 and 5 and Plate 2). This mimics the present structural situation west of segments 1a and 1b, described above (Plate 1 and Figure 6). A similar geometry probably also occurs west of the other stalled COB segment (northeast of Misima Island) given that east and west Misima comprise low-grade above high-grade metamorphic core complexes, respectively, separated by a WNW striking mylonitic ductile shear zone, and inferred to be middle and lower crust unroofed in the footwall of a major offshore detachment [Hill, 1995]. From these three examples it appears that low-angle normal detachments occur in areas of focused strain at and/or before times when the migration of an adjacent spreading segment stalls. Whether such strain focusing and low-angle faulting is cause or effect of the stalled spreading migration is unclear. Nor can we ascertain without more examples whether this temporal/spatial strain association is a necessary and/or suf-
Figure 6. Stacked, migrated, and depth converted 196-channel (EW95-1369) and 148-channel (EW95-1374) seismic sections collected with a 5-km streamer across the rift basin north of Moresby Seamount (Plate 1) that is the locus of current deformation ahead of spreading segment 1 (the western tip of the neovolcanic zone is hatched in the inset). There is no vertical exaggeration. The bounding low-angle normal detachment wraps around Moresby Seamount and has a true dip of 27°±3° towards 015°. Structure contours from 3 to 9 km depth are shown in the inset, with bathymetric contours labeled in hundreds of meters. The antithetic hanging wall normal fault dips south at 45°. On line EW95-1369 the planar detachment (curvilinear in three dimensions) is imaged over the full depth extent of the seismogenic zone (3-9 km) determined from earthquake waveform inversion results [Abers et al., 1997]. Beneath the rift onset unconformity (ROU), Miocene strata on the southern flank of a prerift basin dip north at ~10° beneath the northern margin.

The sufficient condition to form low-angle normal detachments in rifting continental margins.

Note that there is no evidence for thermal uplift or igneous underplating of the Moresby Transform continental margins as spreading axes pass by them, including where the robust segment 2 axis abuts the southern transform COB segment today (Plate 1 and Figure 4). Profiles of bathymetry along and across the margins adjacent the ridge intersections, and of crustal thickness inferred from three-dimensional (3-D) gravity modeling (F. Martinez, unpublished data, 1998), show no evidence for current or previous perturbations of margin depth or crustal thickness by the adjacent ridge tips. This is not a surprising result given that such effects are not generally observed at oceanic ridge-transform intersections. It is in sharp contrast, however, to proposed explanations for thermal uplifts and marginal ridges of several kilometers height, and igneous underplating of 10-km-thick layers, on continental transform margins such as the southern Exmouth Plateau [Lorenzo et al., 1991] and equatorial west Africa [Basile et al., 1993; Maslare et al., 1995]. Note that ODP Leg 159 drilling results did not confirm these hypotheses: transform tectonics, rather than thermal uplift from contact with a hot oceanic accretion center, were responsible for the tilting and uplift of the equatorial west African marginal ridge [Basile et al., 1998].

An evolution similar to segments 2 and 3 west is inferred for the poorly surveyed boundaries of segments 3 east and 5. Segment 5 nucleated from 155°E to 156.5°E, slightly before the eastern half of segment 3 nucleated about 40 km farther south. In this case, when the 155°10'E transform joined the two segments, it truncated the western end of segment 5 and spawned the eastward propagating segment 4 (Plate 1). If this inferred reconstruction is correct, then none of the offsets between spreading segments 1 through 5 initiated as transform faults. They all began as overlapping and/or propagating nontransform offsets (e.g., Plate 2) and the spreading center offsets west of 154°E continue in that mode (Plate 1). The Moresby and 155°10'E Transforms initiated by crosscutting previous rift structures to link spreading segments that had nucleated in, and/or propagated into, offset continental rifts (e.g., Plate 2). They are not directly associated with, nor did they evolve from, obvious transverse structures (transfer/transform faults or aulocogens) in the rifted margins [Taylor et al., 1995].

Spreading is poised to propagate westward from segment 1a along the rift basins north of Moresby Seamount and Normanby Island. West of 151°E, however, earthquake and multichannel seismic (MCS) data show that current extension is not centered on the metamorphic core complexes of the D'Entrecasteaux Islands but is offset more than 40 km to the
south in Goodenough Basin (Plate 1) [Abers, 1991; Taylor et al., 1995; Mutter et al., 1996; Abers et al., 1997]. If extension remains localized there, a spreading segment will nucleate within the 130-km-long Goodenough rift basin and the cycle of a nontransform offset evolving to a transform fault likely will be repeated.

7. Synchronous Rift Onset

Although the history of seafloor spreading in the Woodlark Basin is well constrained by the magnetic lineations and seafloor fabric (Plate 1 and Figure 1), the detailed temporal evolution of rift basin development is not known regionally. Thus, although we know the sequential timing and location of breakup and initial seafloor spreading, only on the northern margin west of 152°E do commercial and academic drilling results and dense grids of MCS lines provide details of the evolution of rifting [Tjhin, 1976; Pinchin and Bembright, 1985; Francis et al., 1987; Taylor et al., 1999]. Furthermore, only in the latter region are the rift basin structures and stratigraphy sufficiently contiguous and well known to be regionally correlated. Farther east, and on much of the southern margin, the rift basins sediments are structurally isolated along and across strike and undated. In this sense, therefore, our perspective is necessarily from the ocean and kinematics, rather than from the continent and dynamics.

Nevertheless, the spreading kinematics have major implications for the history of continental rifting and breakup (see sections 8, 9, and 10), especially when combined with one critical result from drilling and seismic stratigraphy. In the area of the Woodlark Rise west of Woodlark Island to 148°E, a regional angular unconformity separates the synrift sedimentary sequence from an older forearc basin and basement [Tjhin, 1976; Pinchin and Bembright, 1985; Francis et al., 1987; Taylor et al., 1999] as can be seen, for example, on Figure 6. The onset of rift subsidence and sedimentation above this regionally correlative unconformity has been biostratigraphically dated in three wells (located on Plate 1) as occurring sometime between 5.54 and 9.63 Ma at Goodenough 1, between 3.75 and 9.63 Ma at Nubiam 1, and, most precisely, as between 5.54 and 8.6 Ma at Ocean Drilling Program (ODP) Site 1115 [Francis et al., 1987; Taylor et al., 1999]. Within the possible age range of 5.54-8.6 Ma, deposition of the earliest sediments, which are paralic, may be eustatically controlled so that the onset of rift subsidence is likely about 6 Ma. Given that the oldest magnetic lineation in the eastern Woodlark Basin, where the Pocklington continental margin is very narrow, is 3A.1 (6 Ma, Plate 1), these dates are consistent with the onset of continental rifting occurring between 148° and 158°E at 6 Ma or perhaps slightly before. In other words, the onset of rifting of the Papuan continent is synchronous over 1000 km to within the resolution of the available dates.

A similar result has been confirmed by stratigraphic and fission track dating of the onset of rifting in the early Oligocene (~34 Ma) in the Red Sea and Gulf of Aden [Hughes and Beydoun, 1992; Omar and Steckler, 1995; Fantozzi, 1996]. Their data and ours support a rigid plate model for continental extension in which deformation in a rift zone initiates simultaneously along the length of a plate boundary. Strain/rift localization and oceanic accretion can propagate within this already extending zone. The alternative model, in which initial rifting is limited to a segment of the future plate boundary, would require internal deformation of the plate as one part is rifted and another remains intact, and this model is not supported [e.g., Courtillot, 1982; Om and Steckler, 1995].

8. Rift and Spreading Segmentation

Spreading center nucleation, as evidenced in the centers of COB segments 1 and 2 and inferred for COB segments 3 east and 5, results in concordance between oceanic isochrons, continental rift structures and COBs. In contrast, spreading propagation, as evidenced in COB segment 3 west and in the extremities of COB segment 2, occurs by lengthening an existing spreading segment and results in discordant oceanic isochrons and COBs. In the former case, the onset of seafloor spreading is “synchronous” for margin segments of order 50 km long. Nucleation can be distinguished from propagation both conceptually and in terms of rates. Spreading propagation rates vary globally from 10 to 1000 mm/yr (10^-2-1 m/yr) [e.g., Cormier, 1997]. Individual dikes injection events from spreading center volcanoes such as Axial Seamount and Krafla occur at 0.1-6 m/s (∼0.3-20 km/h or ∼10^2-10^3 m/yr) [Elmarsho and Brandsdottir, 1980; Dziak et al., 1995]. However, the full 1975-1984 Krafla tectonomagnetic event rifted and intruded the 80-km spreading segment over 9 years, giving effective lateral rates of order 10^6 m/yr [Tryggvason, 1984]. If similar rates (several orders of magnitude faster than existing spreading segments propagate) characterize the nucleation of a new spreading segment, then the latter may be considered synchronous at geologic time scales of several hundred thousand years.

We observe spreading nucleation to have occurred in the center of late stage rift basins that are separated along strike by accommodation zones (e.g., segment 2, Plate 2, and the failed spreading segment south of Woodlark Island, Plate 1). However, given that spreading segments, once nucleated, usually grow farther by propagation, there need he no direct correlation between seafloor spreading segmentation and rift segmentation [Bosworth, 1986; Rosenblatt, 1987], despite a considerable literature that suggests otherwise [e.g., Lister et al., 1986; Manighetti et al., 1997]. This is the case in the Woodlark Basin, where the oceanic transform and nontransform offsets do not reflect transverse rift structures in the conjugate margins but principally developed at or after breakup (Plate 1) [Taylor et al., 1995].

Furthermore, even the fundamental architecture of rifting, including the location of accommodation zones, can change through time. A good example is the juxtaposition of the accommodation zone at 9°50', 153°15' with the center of the graben immediately north of it and the associated switch to south dipping, high-angle normal faults from north dipping, low-angle detachments (Figures 3 and 5 and Plate 2). The location of this accommodation zone does correlate spatially with the change in the initial propagation of spreading segment 2 from rapid to slow and controls the eastern limit of the failed spreading segment to the west. Within 1 Myr, however, continued westward propagation of spreading segment 2 and the nucleation of spreading segment 1 removed the correlation between the spreading and rifting segmentation.
Figure 7. Plot of cumulative continental and oceanic widths of the Woodlark Basin versus (bottom) longitude with (top) calculated prerifting and stretched continental lithosphere and corresponding strain and strain rates. The diagonal isochrons, labeled in Ma, are derived from the best fitting poles of opening and are dashed where extrapolated from 3.6 Ma to 6 Ma. Total opening is partitioned between continental stretching and seafloor spreading. The width of continental lithosphere includes prerifting and stretched components. The limits of oceanic lithosphere plot parallel to isochrons for nucleated spreading segments, at high angles to isochrons for continent-ocean boundaries formed by slowly propagating spreading segments, and at constant longitude for stalled segments. Continental breakup occurred after 200±40 km of stretching and 130-300% strain and has migrated westward at an average rate of 142 mm/yr since 3.6 Ma. The full history prior to 3.6 Ma, and a record of any spreading prior to 6 Ma, have been lost to subduction eastward beneath the Solomon Islands (Plate 1). The local spike in the strain and strain rate curve is probably the result of subduction crosion forming the notch in the prerifting continental width near the trench-transform intersection at 153.5°E.

9. Strain Rates

The progressive westward migration of Papuan continental breakup occurred in a stair-step fashion [Taylor, 1987]. This can be seen in Figure 7, which plots the width of continental and oceanic lithosphere as a function of longitude, with superimposed isochrons depicting the amount of total opening (riifting plus spreading). The total opening is calculated from the relative motion poles derived from the history of seafloor spreading and assumes that the continental frag-
ments and adjoining oceanic lithosphere outside the extending rift/spreading zone acted as rigid plates (as discussed in section 7). The 0.52 3.6 Ma pole is extrapolated back to 6 Ma (the oldest age of unsubducted oceanic crust in the easternmost basin and the approximate age of rift onset defined by well data west of 152°E, Plate 1) to derive the dashed isochrons. For COBs formed by spreading center nucleation the limit of oceanic lithosphere plots parallel to the isochrons. For COBs formed by fast and slow spreading center propagation this limit plots at small and high angles to the isochrons, respectively. Continental breakup migrated west from Simbo Transport (136.2°E) at an average rate of 14 cm/yr but did so discontinuously [Taylor, 1987], with periods of spreading center nucleation, propagation, and stalling (Figure 7).

There is a rather restricted range of calculated continental stretching (200x40 km) prior to breakup (Figure 7). Within this range, presumably local heterogeneities and/or varying strain histories determined when seafloor spreading initiated. The values are lowest in the east where the continental crust of the Woodlark and Pocklington Rises is narrower and gravity data [Goodliffe et al., 1999] suggest that it is thinner. The continental stretching has a serrated pattern complementary to the amount of seafloor spreading: it increases westward along segments formed by slow propagation and decreases toward the opening pole along nucleated or rapidly propagated segments (where breakup followed localized rifting).

The average strains (total continental width divided by perifit continental width) that characterize the margins at breakup vary by a factor of ~3; minimum values vary from 130% to over 300% if calculated with the 0.52-3.6 Ma opening pole extrapolated back to 6 Ma (Figure 7). Associated strain rates, averaged across the full margin widths and the time from 6 Ma until breakup, are of the order 10^-14 s^-1. Because rifting was spatially and temporally variable, local strain rates may have been 1 or 2 orders of magnitude faster, especially immediately preceding breakup. To a first order, the average continental strain rates increase eastward (i.e., with increasing opening rates), in accord with global compilations of regional strain rates versus displacement rates on faults [Nicol et al., 1997].

10. Comparison With Other Models

These observations are in accord with some, but not all, of the continental breakup models derived primarily from observations in NE Africa (Red Sea, East African Rifts, Gulf of Aden). Our data agree with the Red Sea model of Martinez and Cochran [1988] and Cochran and Martinez [1988] that accommodation zones focus spreading nucleation to the centers of late stage rift basins where normal faulting and crustal thinning have been maximized. In more volcanic margins, such strain focusing could also occur between deeply sourced rift magmatic centers, such as associated with hot spots and volcanic arcs [Taylor et al., 1991; Taylor, 1997; Parsons et al., 1998]. We do not require punctiform initiation of spreading cells above upwelling Rayleigh-Taylor asthenospheric instabilities [Bonatti, 1985]. Nor do we observe the shortening of rift basins as continents approach breakup that Hayward and Ebinger [1996] and Ebinger and Hayward [1996] postulate characterizes the East African Rifts and Afar. For example, the late stage rift basin within which segment 2 nucleated was 140 km long: longer than adjacent rift basins farther from the site of eventual breakup (Plate 2). We do observe the primary rheological control that relatively weak continental versus strong oceanic lithosphere places on the location of rifting [Vink et al., 1984; Steckler and ten Brink, 1986]. Continued nucleation along, or propagation into, continental riffs that approached the oceanic lithosphere of the Solomon or Coral Seas was not favored; east of 154°E the breakup stepped south and then, farther west, stepped north in order to stay centered within the SW trending, then NNW trending, continental domain.

The Woodlark Basin evolution is similar to elements of the model envisioned for the Gulf of Aden by Manighetti et al. [1997]: successive phases of tilting localization, spreading center nucleation, spreading center propagation, and then a jump to the next offset rift. We note, however, that in the Gulf of Aden between 45° and 55°E the magnetic [Cochran, 1981; Tamsett and Searle, 1990] and stratigraphic [Akhmet et al., 1986; Bott et al., 1992; Hughes and Beydoun, 1992; Fantazz', 1996] data are more consistent with the synchronous nucleation of multiple left-stepping rift (34 Ma) and spreading (12 Ma) segments, rather than with the rift and spreading center propagation model proposed by Manighetti et al. [1997] (note especially Figure 17 of Fantazz' [1996]). Thereafter, a prolonged stall (~12-5 Ma) occurred at the Shukra-El Sheik discontinuity (45°E: M fracture zone of Cochran [1981]) before spreading penetrated the Gulf of Tadjura and lithosphere affected by the Ethiopian hot spot.

We observe elements of the three models of rift propagation into continents proposed by Vink [1982], Courtillot [1982], and Martin [1984], though none of these individual models fully matches the Woodlark Basin data. As in Vink's [1982] model, the New Guinea continental lithosphere is locally nonrigid around the breakup zone and, prior to seafloor spreading, stretches in proportion to the amount of seafloor spreading farther from the pole of opening. As in Courtillot's [1982] model, there are both spreading segments that nucleate and others that slowly propagate through zones of distributed deformation [McKenzie, 1986]. As in Martin's [1984] model, the pole of opening is nearby, and the amount of continental stretching prior to spreading initiation varies along strike, so that some spreading segments nucleate ahead of a propagator.

These model elements, together with stalled segments, rheologically controlled offsets, and oceanic transform development, are combined in our preferred model of continental breakup shown in Figure 8 that conceptually depicts the central Woodlark Basin in four stages from 4 Ma to 1 Ma. Additional aspects of continental breakup observed in the Woodlark Basin evolution to the present, including continued extension of the margins for up to 1 Myr after spreading initiates which sometimes results in abandonment of an initial spreading segment and nucleation of another (Plate 1) [Taylor et al., 1995], are not depicted in this already complex model.

11. Conclusions

Rifting along the length of the prototypical present Papuan Peninsula of continental New Guinea spawned seafloor spreading in the Woodlark Basin since 6 Ma. The continental breakup and westward migration of the spreading center, averaging 14 cm/yr since 3.6 Ma, was progressive but
discontinuous, with periods of spreading center nucleation, propagation, and stalling. Spreading center nucleation results when upwelling asthenosphere feeds magmas into elongate sites of focused continental extension/thinning and produces isochronous continent/ocean boundary (COB) segments that are discordant with rift structures and seafloor isochrons. Spreading propagation lengthens an already nucleated segment and results in COBs that are discordant with oceanic isochrons and may be either discordant or discordant with rift structures on the margins, depending on whether or not the regional rifting is localized in front of the propagator. Stalling of the spreading propagation, or development of a ridge-ridge transform fault, results in oceanic isochrons that are close to perpendicular to the COR.

The fundamental architecture of rifting, including the location of accommodation zones and the mode of extension, can change through time. Accommodation zones focus spreading nucleation to the centers of late stage rift basins where normal faulting and crustal thinning have been maximized. We document major low-angle normal detachments as well as high angle normal faults in the rifted continental margins. Both modes of extension are active at different times and places within the rifting continent, and either mode may focus breakup.

Continental breakup is focused currently on an asymmetric rift basin bounded by a low-angle (27°-33°), curviplanar, normal detachment that we image over the full depth extent of the seismogenic zone (3-9 km) determined from earth-
quake waveform inversion results by Abers et al. [1997]. It is one of three examples in the Woodlark Basin where low-angle normal detachments occur in areas of localized strain at, and/or before, times when the migration of an adjacent spreading segment stalls. In contrast, breakup at 2 Ma occurred along a symmetric rift basin bounded by high-angle normal faults.

The continental breakup proceeds by successive phases of rifting localization, spreading center nucleation, spreading center propagation (sometimes stalling), and then a jump to the next site of localized rifting. Extension of the margins may continue for up to 1 Myr after spreadinginitiates, which sometimes results in abandonment of an initial spreading segment and nucleation of another [Taylor et al., 1995].

We have estimated the rates of continental rifting and the total continental stretching from the basin spreading kinematics. The range of calculated continental extension is 200±40 km prior to breakup, and the average continental strain varies from 130 to 300%. Associated average continental strain rates are of the order 10^-14 s^-1. Locally and over shorter periods of time, strain rates may have been 1 or 2 orders of magnitude faster, especially during localized rifting immediately preceding breakup.

Margin offsets are rheologically controlled to first order by the geometry and weakness of the Papuan continental lithosphere (relative to the stronger oceanic lithosphere of the surrounding Solomon and Coral Seas) and only secondarily by stress concentrations at the tip of propagators. The westward migrating breakup stepped south, then north, to stay centered within the continental domain.

Spreading propagation and transform development have modified any initial correlation between seaﬂoor spreading segmentation and margin rift segmentation. The initial spreading center offsets, formed by stepwise spreading nucleation, developed overlapping and/or propagating nontransform offsets. The Moresby and 155°10'W transforms initiated by crosscutting previous rift structures to link spreading segments that had nucleated in, and/or propagated into, offset continental rifts. They did not evolve from transform/transform faults or autogenous in the rifted margins.

There is no evidence for thermal uplift or igneous underplating of the Moresby Transform continental margins as spreading axes pass by them, including where spreading segments 2 and 3 currently intersect the transform COB segments.

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