

The boundary between continental rifting and sea-floor spreading in the Woodlark Basin, Papua New Guinea

A. M. GOODLIFFE¹ & B. TAYLOR²

¹*Department of Geological Sciences, University of Alabama, Tuscaloosa, Alabama, USA (e-mail: amg@ua.edu)*

²*School of Ocean and Earth Science and Technology, University of Hawaii, Honolulu, Hawaii, USA*

Abstract: Seismic reflection, bathymetry, acoustic imagery and magnetic data are presented that encompass the boundary between rifting of the Papuan continent and westward propagating sea-floor spreading in the Woodlark Basin. West of the spreading tip, the southern margin is characterized by large fault blocks, which were tilted to the south on north-dipping normal faults during the current rifting phase, and graben further south where previous rifting failed. The northern margin is devoid of large offset normal faults and its subsidence from near sea level requires synrift flow of the lower crust. The margin asymmetry primarily reflects across-strike differences in the prerift geology, morphology and rheology. Horst and graben with 3 km relief below sea level occur adjacent to the spreading tip. Although heat flow data imply greater crustal thinning to the south of Moresby continental seamount, seismicity shows that deformation is currently focused on the Moresby normal fault that bounds its northern side. Based on logging results from ODP Leg 180 and semblance analysis, depth conversion of seismic reflection data show that Moresby Fault has a dip $\geq 29^\circ$, which is compatible with an earthquake slip plane of 30° – 33° . Directly east of Moresby Seamount, the first spreading segment is identified on the basis of magnetic, bathymetry and sidescan data to comprise two ring dykes in the west and Cheshire Seamount in the east. A seismic reflection line crossing the western ring dyke shows an intrusive body that cuts Moresby Fault at depth. Currently, intrusive rocks have only reached the surface near the ring dyke, leaving flat-lying synrift sediments, sills (?) and surface lava flows overlying the main intrusive body. The boundary between continental rifting and sea-floor spreading in the Woodlark Basin is spatially abrupt, although the processes overlap in time. The progression from rifting to spreading is characterized by a decrease in sedimentation as the margins are progressively thinned, subside below sea level, are eroded less and trap sediments in proximal basins. The post-rift sedimentation near the continent–ocean boundary is hemipelagic and drapes breakup topography without a distinct breakup unconformity. Synrift sediments, deposited above the 8.4 Ma rift onset unconformity and prior to breakup, are characterized by rotated sections (with parallel or diverging reflectors) as well as by ponded sections that onlap bounding fault blocks where local extension has ceased. Splayed reflectors representing sediment deposition on rotating fault blocks do not characterize the full synrift section, but only those times and places where extension is localized.

The transition between continental rifting and sea-floor spreading marks the boundary between two fundamentally different modes of strain accommodation. How this transition takes place is poorly understood at all scales. Much of our present knowledge is based on observations from old and cold passive margins where the location of the conjugate margin is ambiguous and the basement is buried by thick sediments or voluminous magmatism (Holbrook & Kelemen 1993; Driscoll & Karner 1998). Active rifting–spreading boundaries are found only in a small number of locations. The Red Sea and the Gulf of Aden, both in cratonic continental settings, are currently forming as a result of rifting of the Arabian Peninsula from Africa (Cochran 1981; Martinez & Cochran 1988). The

Woodlark Basin and the Gulf of California (Moore 1973; Taylor *et al.* 1995, 1999; Fletcher & Munguia 2000) are also the result of continental breakups, but the tectonic history is made more complex by the interaction with recently active subduction systems.

Several kinematic models have been proposed to explain basin-scale variations in features such as continental margin width, breakup style and spreading centre segmentation by varying the amount of extension prior to the initiation of sea-floor spreading (Courtillot 1982; Vink 1982; Martin 1984). Finite-element models have been used to match the observed data by varying the thickness and composition of the crust, the geotherm, the extension rate, and the strength and distribution of pre-existing

weak zones (Bassi *et al.* 1993; Dunbar & Sawyer 1996; Harry & Bowling 1999; Lavier & Buck 2002; Corti *et al.* 2003; Van Wijk & Blackman 2005). Modelling studies have been hampered by a lack of accurate kinematic constraints on parameters such as strain rate, duration of rifting preceding rupture, and prerifting continental width and crustal thickness.

The evolution of rift architecture through time is typically described as a transition from broadly distributed extension, where large crustal blocks rotate along faults that sole out at depth or transition to ductile deformation, to a narrow zone of focused extension that allows rapid lithospheric thinning and melt generation (Cochran 2005). In contrast, South Atlantic passive continental margins (Karner *et al.* 2003) are characterized by the stacking of regional synrift sag basins. Large, late-stage rift faults in the South Atlantic margins do not control the regional development of synrift accommodation and are thus a small part of the extension process.

Late-stage subsidence during the rifting–spreading transition is usually explained by interpreting the margin being studied as the upper plate in a simple shear model. This has resulted in the ‘upper plate paradox’ (Driscoll & Karner 1998), where both conjugate margins have been interpreted as the upper plate. This paradox can be avoided by using lower crustal and upper mantle deformation to create subsidence during late-stage continental extension (Karner *et al.* 2003). Subsidence may also originate from cooling and contraction of anomalously hot asthenosphere in the region of deformation (Buck 2004) or be the result of an upwelling divergent flow field within continental lithosphere and asthenosphere (Kusznir *et al.* 2004).

The processes involved in the initiation and growth of a spreading centre in a continental rift have been examined on a large scale at several locations. In the Red Sea, continuous sea-floor spreading in a well-defined axis in the south becomes discontinuous to the north, forming discrete deeps interpreted as cells of sea-floor spreading (Bonatti 1985). The regular spacing of the deeps and the similar spacing of discontinuities in the southern Red Sea suggest that both may be caused by asthenospheric upwelling, the wavelength of which is controlled by the ratio of the viscosities and the layer thicknesses. Once a diapir reaches the surface at the centre of a segment it continues to grow along strike, eventually merging with its neighbours to create a continuous spreading centre. During rifting, adjacent segments may be offset by accommodation zones that eventually form the site of oceanic transform faults (Cochran & Martinez 1988). In the Woodlark Basin, two distinct modes of sea-floor spreading initiation have been identified. One involves propagation of sea-floor spreading into the rifting continent; the other nearly simultaneous nucleation of

sea-floor spreading along more than 100 km-long rift segments (Taylor *et al.* 1999). The former results in a continent–ocean boundary (COB) that is highly oblique to early margin structures, abyssal hill fabric and magnetic isochrons. In the later case the COB, abyssal hill fabric and magnetic isochrons are all parallel. Other studies of the geological and dynamic processes by which sea-floor spreading initiates in a continent (Rosendahl 1987; Hayward & Ebinger 1996) demonstrate that many of the details are highly debated. One shortcoming of all these studies is that they have not determined how the spreading centre first ruptures the surface and how this location is related to the late-stage continental faulting.

The Woodlark Basin is perhaps the only place where the COB has been located regionally at a kilometre scale. Sediment cover is generally thin, and the opening kinematics, strain history and margin subsidence have been intensely investigated (Taylor *et al.* 1995, 1999; Goodliffe 1998; Martinez *et al.* 1999; Taylor & Huchon 2002). This affords us an unprecedented view of the detailed processes involved in: (1) continental rifting directly preceding the initiation of a spreading centre; (2) the mechanics of spreading centre emplacement; and (3) the early evolution of a spreading centre.

Spreading history and Euler poles

Rifting of the Papuan Peninsula started at approximately 8.4 Ma (Taylor & Huchon 2002), and led to separation of the once contiguous Woodlark and Pocklington rises that bound the Woodlark Basin (Fig. 1). The oldest oceanic lithosphere in the Woodlark Basin, formed during magnetic chron 3A.1 (Taylor *et al.* 1999) (*c.* 6 Ma; Cande & Kent 1995), is found at the eastern end of the basin adjacent to the southern margin; any older oceanic lithosphere has been subducted under the Solomon Islands. Reconstructions based largely on fracture zone trends (Goodliffe 1998) show that since magnetic chron 3A.1 the majority of basin opening can be described as a rotation about a single Euler pole ($147^\circ + 1^\circ - 2^\circ\text{E}$, $9.3^\circ\text{S} \pm 0.2^\circ\text{S}$, $4.234^\circ\text{Ma}^{-1}$ Fig. 2), implying that rifting initiated simultaneously throughout the basin. Abyssal hill fabric, magnetic isochrons, a major low-angle fault bordering Misima Island (Hill 1995), faults bounding Milne Bay graben (Jongsma 1972) and Goodenough Fault, a structure bounding the southern side of Goodenough Basin (Fang 2000) (Figs 1 & 2), are all co-polar with this Euler Pole.

Sea-floor spreading propagated in a stair-step fashion from the east into rifting crust at an average rate of approximately $140\text{ km per Ma}^{-1}$ (Taylor & Exon 1987), reaching Moresby Seamount, where the rifting to spreading transition is

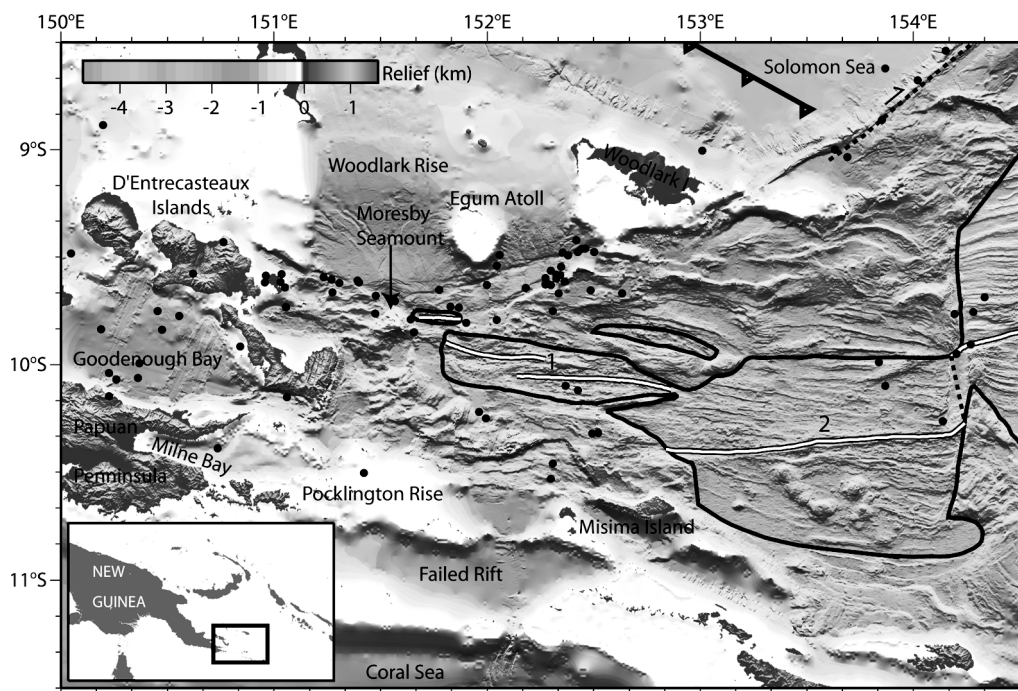


Fig. 1. Regional setting of the boundary between continental rifting and sea-floor spreading in the Woodlark Basin. The conjugate Woodlark and Pocklington Rises have separated over the last 8.4 Ma as a result of rifting the Papuan Peninsula and the westward propagation of sea-floor spreading (accreted oceanic crust is outlined by a solid black line; the spreading centre is depicted by a black line with white centre). Dashed black lines indicate active strike-slip motion. Teleseismicity is shown by black circles.

currently found. The 500 km-long active spreading centre, located on the basis of bathymetry and acoustic imagery, comprised five main segments at the start of the Brunhes chron (0.78 Ma). An up to 22° counter-clockwise reorientation of the spreading centre at 80 ka (Goodliffe *et al.* 1997) in response to a shift in the Euler pole at 0.52 Ma (Goodliffe 1998) (Fig. 2) resulted in the subdivision of many of the spreading segments into shorter segments that are now oblique to the Brunhes isochron. The modern Euler pole (12°S, 144°E, $2.437^\circ \text{ Ma}^{-1}$; Fig. 2) is defined largely by the offset of pre-existing fracture zones and is close to the Australia–Woodlark Euler pole (11.28°S, 147.62°E, $2.82 \pm 0.18^\circ \text{ Ma}^{-1}$) (Wallace *et al.* 2004) determined by simultaneously inverting GPS data and earthquake slip vectors with transform orientations.

Spreading segment 1 extends from the current spreading tip at 151°40'E–152°51.6'E and comprises three subsegments, 1a, b and c (Fig. 2). Sea-floor spreading on this segment nucleated between 151°45'E and 152°51.6'E at approximately 0.8 Ma (Taylor *et al.* 1995; Goodliffe 1998). At the resolution of the available data and analysis techniques the COB is coincident with the Brunhes isochron

and parallel to great circles about the pre-0.52 Ma Euler pole (Fig. 2). The change in Euler pole resulted in the subdivision of segment 1 into two shorter segments and an approximately 8° rotation. Unlike segments 2–5, segments 1b and 1c are closer to co-polar with the pre-0.52 Ma Euler pole and the Brunhes isochron than to the current pole.

To the north of segment 1, the seismically active Egum Graben (Figs 1 & 2) cross-cuts normal faults in the northern margin of segment 1 and has a trend that is co-polar with the present Euler pole. Although this may be coincidental, it may mark the reactivation of a favourably trending structure by the change in Euler pole. This might also explain the concurrent sea-floor spreading and continental rifting at this longitude (Taylor *et al.* 1995). The fault bounding Egum Graben has been interpreted as a continuation of the strike-slip fault separating the eastern Woodlark Basin from the Solomon Sea (Taylor *et al.* 1999). This transtensional fault system may bound a third (Trobriand) plate, including the margin to the north of Moresby and Egum Graben, Egum Atoll and Woodlark Island (Davies *et al.* 1984), and form a triple junction with

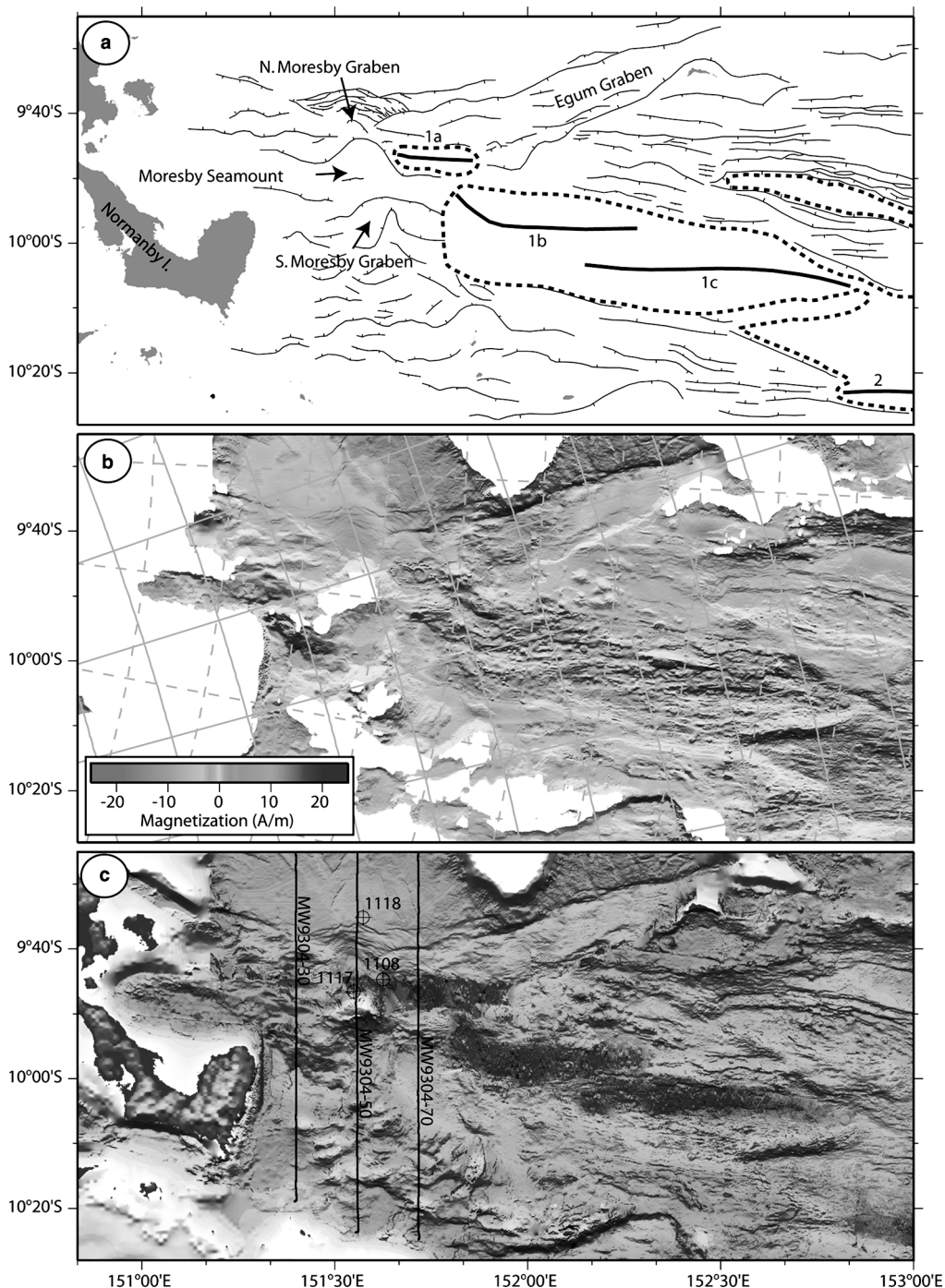


Fig. 2. Map of major faults seen at the surface in the vicinity of the rifting–spreading transition (a). The axes of the active neovolcanic zones are shown by thick solid lines and the limit of oceanic crust is dashed. Sea-floor magnetization illuminated from the north by bathymetry (b). Oceanic crust formed since the start of the Brunhes chron (0.78–0 Ma) on segment 1 has a positive magnetization. At the resolution of the data the continent–ocean boundary and the 0.78 Ma isochron are coincident. Small and great circles about the present Euler pole (12°S , 144°E , $2.437^{\circ}\text{Ma}^{-1}$, solid grey lines) and the pre-0.52 Ma Euler pole ($147^{\circ} + 1^{\circ}\text{--}2^{\circ}\text{E}$, $9.3^{\circ}\text{S} \pm 0.2^{\circ}\text{S}$, $4.234^{\circ}\text{Ma}^{-1}$,

spreading segment 1a and the Papuan rifts. Given the complex tectonic setting of segment 1a it is not clear that it is opening around the same Euler pole as the spreading segments further east.

The continental margins proximal to the initiation of sea-floor spreading

Using swath bathymetry data and three representative six-channel seismic lines (see Appendix 1 for details of the data used and processing steps), we first describe the margin structures that surround sea-floor spreading segment 1. The margin of the Woodlark Basin south of segment 1 and Moresby Seamount (Fig. 2) is characterized by large fault blocks, trending on average east–west but locally NE–SW, that are typically tilted to the south and overlain by isolated sedimentary basins that record varying degrees of rotations (Fang 2000) (Fig. 3). Poor basin interconnectivity and an absence of sampling make attempts at dating fault motion purely speculative. A scarcity of seismicity in this region (Abers *et al.* 1997; Ferris *et al.* 2002) (Fig. 1) indicates that there is currently little fault motion. An analogy has been made to oceanic rift propagation, with many of the fault blocks rotated to the NE about vertical axes by shear between laterally offset centres of extension (Mutter *et al.* 1996).

Six-channel seismic (SCS) line MW9304-30 (Figs 2 & 3) crosses the current focus of deformation to the west of Moresby Seamount where sea-floor spreading has not yet initiated. At the southern end of the line there are isolated sediment ponds that record tilting to the south of at least two large crustal blocks. The northward tilting of an adjacent progradational sequence illustrates that deformation is locally more complex. A graben centred on SP900 that marks the deepest point at this longitude widens to the east to become the South Moresby Graben (SMG). Further north a series of three fault blocks tilted to the south on north-dipping normal faults sit in the westward extension of Moresby Seamount and North Moresby Graben (NMG). Some complex basement geometries (for example, at SP1560) suggest locally more complex deformation, possibly involving unresolved south-dipping faults. A fourth ridge (SP1900) is an uplifted horst block with a thick synrift sedimentary sequence. To the north, the largely undeformed northern margin with its thick synrift sedimentary sequence dips to the south.

The stratigraphy here is contiguous with that imaged to the east and sampled during ODP Leg 180 (Taylor & Huchon 2002). North of SP2500, prominent intrabased reflectors below the synrift sediments dip to the north. These are believed to originate from sills within the approximately 66 Ma diabase and metadiabase basement (Brooks & Tegner 2001; Monteleone *et al.* 2001).

SCS line MW9304-50, 18.5 km to the east, crosses the SMG, Moresby Seamount, and NMG. Owing to the near coincidence with this line of multichannel seismic (MCS) line EW9510-1366 and the ODP Leg 180 drilling transect, the stratigraphy and basin history of Moresby Seamount and the northern margin is very well known at this longitude. In summary, this involved gradual subsidence from sea level and tilting to the south of the northern margin, with early focusing of extension to the north and south of Moresby Seamount and its subsequent isostatic rise between late bounding faults (Taylor & Huchon 2002) (Fig. 4). It is important to note that, although the northern margin is bounded in the south by the major north-dipping Moresby Fault, the wavelength of subsidence is too great to be due to that fault alone. The absence of large-offset normal faults on the northern margin means that other mechanisms such as flow of the lower crust must be called on to explain its subsidence. The ODP drilling transect did not extend to the southern section of SCS line MW9304-50, leaving the details of the evolution of this part of the basin unexplored.

SMG (Fig. 2) is deepest directly to the south of the peak of Moresby Seamount. Here it is subdivided into two parts separated by an north–south trending ridge that plunges to the north. To the east of the ridge the graben floor reaches 3150 m. Low-amplitude positive magnetization implies that rifting may have been accompanied by magmatism extending west from spreading segment 1b. The graben floor is subhorizontal and sediment covered, with the exception of localized regions of high acoustic backscatter in the north indicative of fault scarps (Fig. 2c). A 4×0.1 km linear feature on the sea floor at $151^{\circ}42'E$, $9^{\circ}57'S$ is interpreted as a lava flow. To the west of the north-plunging ridge the graben floor is characterized by generally smooth sea floor reaching depths of 3000 m. A somewhat rougher sea floor close to Moresby Seamount is likely to be the result of debris flows. Consistently high values of heat

Fig. 2. (Continued) dashed grey) are shown. The 0.78 Ma boundary is co-polar with great circles about the pre-0.52 Ma Euler pole, whereas Egum graben is co-polar with great circles about the current Euler pole. Bathymetry illuminated from the north and overlain by acoustic imagery (c) shows the location of the neovolcanic zone (black = high acoustic backscatter) relative to the main fault trends in the vicinity of the rifting–spreading transition. The location of the six-channel seismic lines described in this paper are shown along with ODP sites 1108, 1117 and 1118.

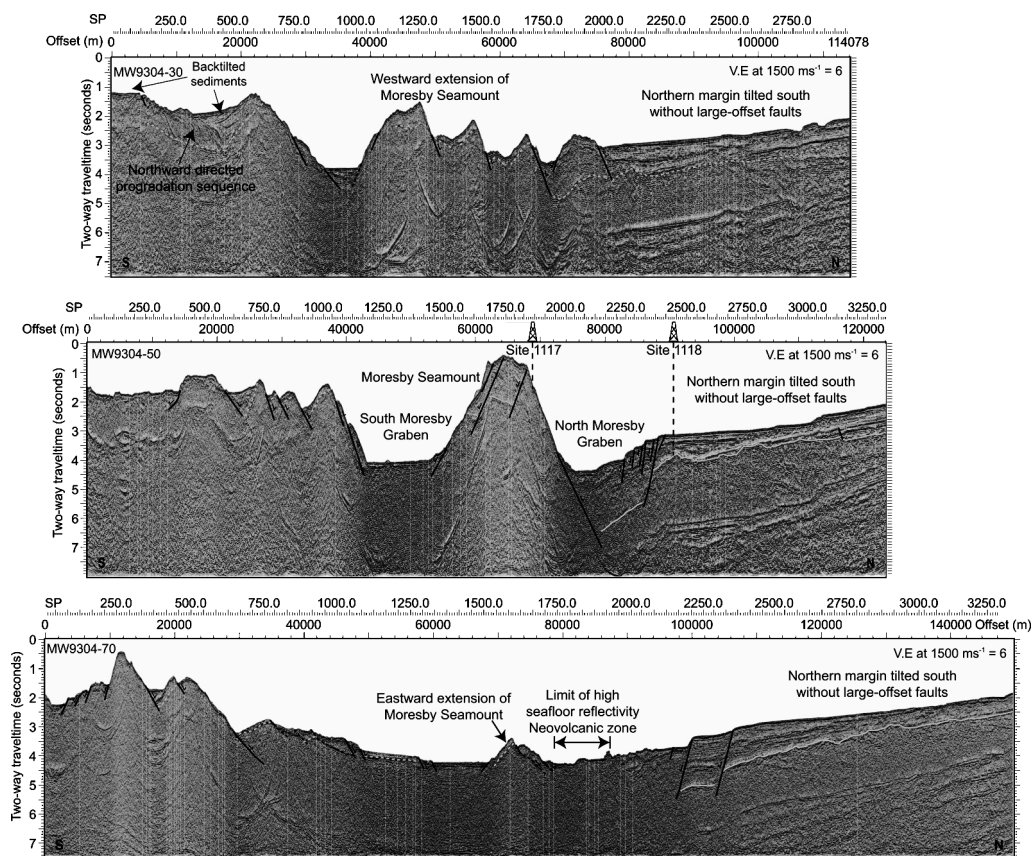


Fig. 3. Six-channel seismic (SCS) lines MW9304-30 (top), MW9304-50 (middle) and MW9304-70 (bottom) across the continental margins in the vicinity of the rifting–spreading transition. The rotation of back-tilted fault blocks is recorded locally by the fanning of sediment packages in half graben that form small isolated basins. The southern margin is dominated by these tilted fault blocks, whereas the northern margin is largely unfaulked, but subsided and south tilted. From SCS line MW9304-30 in the west to MW9304-70 in the east there is a focusing of extension first at North and South Moresby Graben and then on spreading segment 1a. The blue line marks the base of the synrift sediments. ODP Site 1118 is projected from its true location 1.5 km to the east. ODP Site 1117 is projected from its true location 0.7 km to the west. The location of the SCS lines is shown in Figure 2.

flow, reaching a peak of 254 mW m^{-2} , were recorded in SMG (Goodliffe *et al.* 2000). This is in contrast to a maximum of 100 mW m^{-2} recorded in NMG. SCS and MCS data in SMG (Fig. 3) do not satisfactorily image basement. Nevertheless, gravity data indicate a similar amount of crustal thinning as NMG (Goodliffe *et al.* 1999b).

To the south, between SP500 and SP1200 on SCS line MW9304-50, north-dipping normal faults bound crustal blocks that have been tilted to the south. Further to the south at SP390 a south-dipping normal fault forms the northern margin of a shallow rift basin, the eastward extension of which is seen in the form of a series of south-dipping faults on SCS line MW9304-70. The southern side of this rift basin was not imaged

owing to its proximity to the reefs that top Pocklington Rise. Further east, between $151^{\circ}50'E$ and $152^{\circ}30'E$, a scarp dipping approximately $30^{\circ}S$ is seen in the bathymetry (Figs 1 & 2). This fault may form the northern boundary of a basin seen to the west of Misima Island. Still further to the south a large ($>160 \text{ km}$ long and *c.* 40 km wide) basin is bounded by reefs (Fig. 1). The absence of seismicity and the location of this basin away from the focus of current activity suggest that it is an early (failed) rift that predates the rifting that led to breakup and the initiation of sea-floor spreading on segment 1.

To the north of SP250 on SCS line MW9304-70 there are at least three large fault blocks that have been tilted to the south by motion on north-dipping

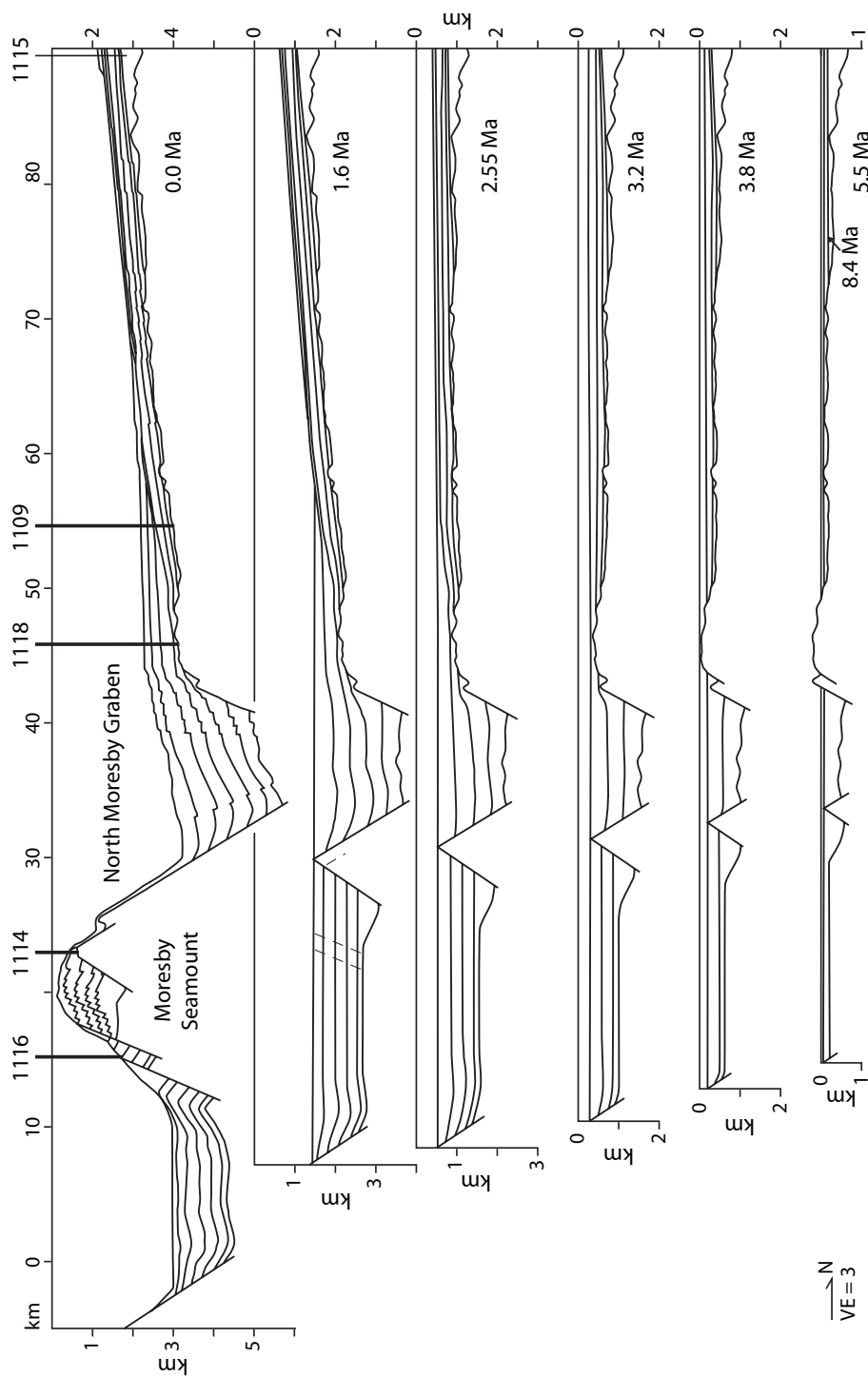


Fig. 4. Interpreted line drawing of MCS line EW9510-1366 (top panel), 1.5 km to the east of SCS line MW9304-50. Reconstructions at 1.6, 2.55, 3.2, 3.8 and 5.5 Ma (lower panels) use constraints from seismic stratigraphy and ODP drilling (magnetobiostratigraphy and palaeowater depths). Adapted from Taylor & Huchon (2002).

normal faults. North of SP800 fault geometries become unclear, possibly because of extreme fault block rotation and subsequent cross-cutting by the late-stage north-dipping faults indicated by the small sea-floor offsets. Locally, the sea floor is underlain by horizontally stratified sediments. This close to the western end of segment 1b, the presence of lava flows, dykes and sills cannot be ruled out. Basement and fault plane reflectors on the small ridge, continuous with Moresby Seamount to the west and centred on SP1600, suggest that it is underlain by south-tilted fault blocks. Directly to the north the sea floor has high acoustic reflectivity and is interpreted to be floored by recent lavas (Fig. 2c). Small ridges on the sea floor at SP1830 and SP1920 are part of an approximately 4 km-diameter circular feature on the sea floor. Below the region of high sea-floor reflectivity there are no coherent reflectors. At SP2195 and SP2350 major south-dipping normal faults down-drop the thick synrift sedimentary sequence of the northern margin to the south. As in SCS lines MW9304-30 and MW9304-50, the margin further north is tilted to the south and shows no evidence of major fault offsets.

The SCS lines, especially MW9304-30 and MW9304-50, show marked margin asymmetry (Fig. 3). The northern margin slopes gradually to the south with little evidence of large-offset normal faults until the NMG is reached. In the south large fault blocks are rotated. At face value this could imply a simple shear origin for the margin (Wernicke 1985; Lister *et al.* 1991), but a detachment fault of sufficient size to explain the wavelength of the subsidence has not been identified. More probably, the structural asymmetry is a function of the pre-existing continental morphology–geology–thickness (forearc basin v. mountainous continental arc and back-arc) combined with lower crustal flow (Goodliffe *et al.* 1999a).

Between 151°55'E and 152°50'E the fabric on the southern and northern rifted margins close to the continent–ocean boundary (COB) is broadly parallel to the N84°W-trending COB (Fig. 2). This is consistent with the initiation of sea-floor spreading almost simultaneously along segment 1 (Taylor *et al.* 1999). The normal faults bounding the margin north of segment 1 are cut by Egum Graben. This more than 100 km-long structure trends N73°E from Moresby Seamount and is the focus of ongoing seismic activity with many $>M_{w6}$ earthquakes (Taylor *et al.* 1995, 1996; Abers *et al.* 1997). To the east the rift narrows and shallows towards the platform surrounding Woodlark Island. At 152°13'E Egum Graben is 17.5 km wide and bounded by scarps with approximately 40° dips. The main rift floor, at a depth of 2500 m, is 600 m below the rift shoulders. A

nested 4.1 km-wide and 50 km-long graben on the southern side of Egum Graben locally reaches a depth of 2600 m and terminates in the west against Cheshire Seamount. There are small edifices to the south of Egum Graben at 152°10'E, 9°50'S that may be volcanic in origin, but are no longer active (Fig. 2c).

At 152°40'E, 10°18'E the southern margin is cross-cut by a WNW-trending COB (Fig. 2). Sea-floor spreading on segments 1 and 2 propagated past one another, forming an overlap that surrounded and rotated counter-clockwise a sliver of continental crust (centred at 152°40'E, 10°10'E) (Goodliffe 1998). Since 80 ka, sea-floor spreading on segment 2 has rotated and jumped to the south. Propagation of the new spreading into the margin has created a small re-entrant at 152°45'E, 10°25'S (Fig. 2c).

Kinematic reconstructions at the longitude of Moresby Seamount (Fang 2000; Taylor & Huchon 2002) of faulting and subsidence north of 10°S show that many of the faults (with dips ranging from 30° to 60°) that are active today were active from at least 5.5 Ma (Fig. 4). Between 5.5 and 1.6 Ma, extension on the major faults only added up to 7 km, compared to more than 12 km between 1.6 Ma and the present day. However, similarities in sedimentation rate between ODP sites 1108, 1109, 1118 and 1114 (Takahashi *et al.* 2002) show that these sites were in a common tectonic setting with no large depositional barriers separating them until at least the late Pliocene (Robertson & Sharp 2002). Although the rift basins in this region were growing, fault growth was being out-paced by sedimentation. The post-1.6 Ma increase in extension rate at the longitude of Moresby Seamount (Taylor & Huchon 2002) and a dramatic decrease in sedimentation on the northern margin at 1.2 Ma (Takahashi *et al.* 2002), implying that the rapid subsidence of NMG and uplift of Moresby Seamount had cut off sedimentation to the north (Sharp & Robertson 2002), both indicate strain localization on the rifts surrounding what would become sea-floor spreading segment 1.

When sea-floor spreading initiated on segment 1 at about 0.8 Ma, the neovolcanic zone extended only from 151°45'E to 152°40'E (Goodliffe 1998). Beyond these limits, strain accommodation and crustal thinning was still not sufficiently focused for spreading to initiate. To the east strain localization was achieved by propagation of the neovolcanic zone, forming a spreading centre that overlapped with segment 2 (Goodliffe 1998). The neovolcanic zone did not propagate to the west, remaining stalled at 151°48'E for approximately 0.7 million years. Strain accommodation on the faults directly to the west of the spreading tip must therefore have still remained distributed, at

least within the confines of the segment 1 rift. This produced the horst and graben terrain that is centred on Moresby Seamount and continues westward to the north of Normanby Island (Figs 1–3). Within this faulted terrain, earthquakes indicate that current deformation is localized on the NMG (Fig. 1), probably at least since the nucleation of spreading on segment 1a about 0.1 Ma.

Faulting directly ahead of the westernmost spreading segment

NMG has been a focus of MCS surveys and ODP Leg 180 drilling aimed at understanding continental breakup and the evolution of a shallow-angle normal fault. We reproduce here just four of the tens of MCS lines across the structure (Figs 5–7). In addition, a line drawing interpretation of MCS line EW9510-1366, which passes through the primary transect of drill sites, is shown in Figure 4 and all of the EW9510 MCS lines are published (Goodliffe *et al.* 1999b).

The MCS data show that NMG is an asymmetric graben bounded to the south by Moresby Fault and to the north by an antithetic normal fault system. The width of the graben between the sea-floor traces of the bounding faults varies from approximately 12 km NNE of Moresby Seamount to more than 20 km on either side, varying as a function of the curvilinear to zig-zag fault patterns (Figs 2 & 5). MCS line EW9510-1369 crosses the NMG near its narrowest (Fig. 5) and images Moresby Fault continuing at depth beneath the antithetic fault and a down-dropped keystone basement block (Fig. 6). Basement drilled on both sides of NMG beneath the rift-onset unconformity (Fig. 4) comprises diabase and, in the footwall of Moresby Fault at ODP Site 1117, gabbro with a crystallization age of 66 ± 1.5 Ma (Brooks & Tegner 2001; Monteleone *et al.* 2001). Heavily faulted Pliocene–Pleistocene synrift sediments cap Moresby Seamount, which rises to almost 150 m bsl (below sea level) (Taylor & Huchon 2002) (Figs 3 & 4). The northern border fault of the NMG is a growth fault, dipping approximately 54° S, with basement offsets of about 1.5 km but surface offsets of only approximately 150 m (Figs 4 & 6). Whereas its growth matched that of Moresby Fault early in the development of the rift, it has been far exceeded in the Quaternary by the faults that bound Moresby Seamount (Fig. 4).

The NMG is contiguous with the trough containing spreading segment 1a. It shares a common northern border fault with Egum Graben, but has a deeper axis. From the western end of the neovolcanic zone the sea floor climbs 150 m over a distance of approximately 5 km before descending to the deepest part of the NMG at 3250 m (Fig. 5) where

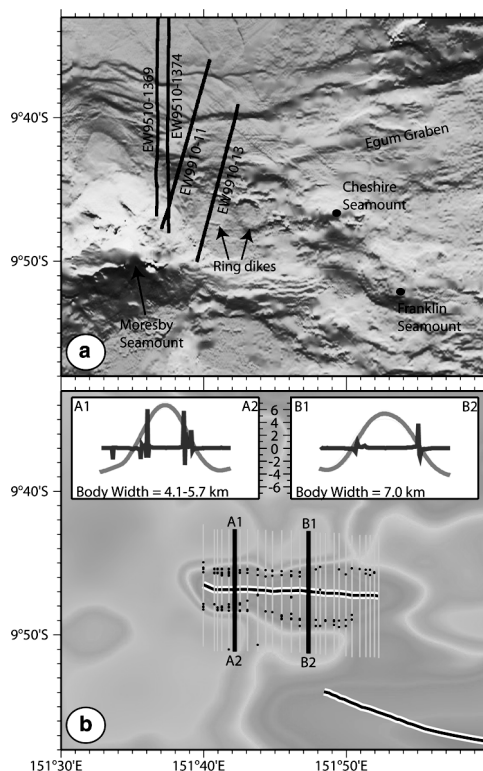


Fig. 5. Bathymetry (a) in the vicinity of the rifting–spreading transition. Key features, including two ring dykes on the seafloor at the western end of segment 1a, are labelled. Cheshire Seamount is a volcanic construct at the eastern end of segment 1a. Sea-floor magnetization (b) for the same area (see Fig. 2 for the colour scale). Profiles along which the horizontal gradient was calculated are shown in grey. Black dots mark the location of gradient maxima. Insets show the magnetization (blue) and horizontal gradient (green) along the profiles A1–A2 and B1–B2. The separation of the maxima is a measure of the width of the causative body assuming a tabular geometry.

the hanging-wall sediments are tilted south and have collapsed the most (Fig. 4). ODP Site 1108 (Fig. 2) was drilled in the bathymetric saddle on MCS line EW9510-1374 (Fig. 6). It and other drill sites at the foot of the Moresby Fault could not reach the fault at depth owing to the presence of coarse, unconsolidated talus shed from the slope.

The synrift sediments north of the NMG are approximately 1 km thick and thin northwards, but those within the graben are more than 2 km thick and subparallel to basement (Figs 4 & 6). The graben basement is almost horizontal on MCS lines EW9510-1369 and EW9510-1374 but dips gently south on MCS lines EW9510-1366 and EW9910-11. Basal onlap is rarely observed. An exception occurs south of SP7200 on MCS

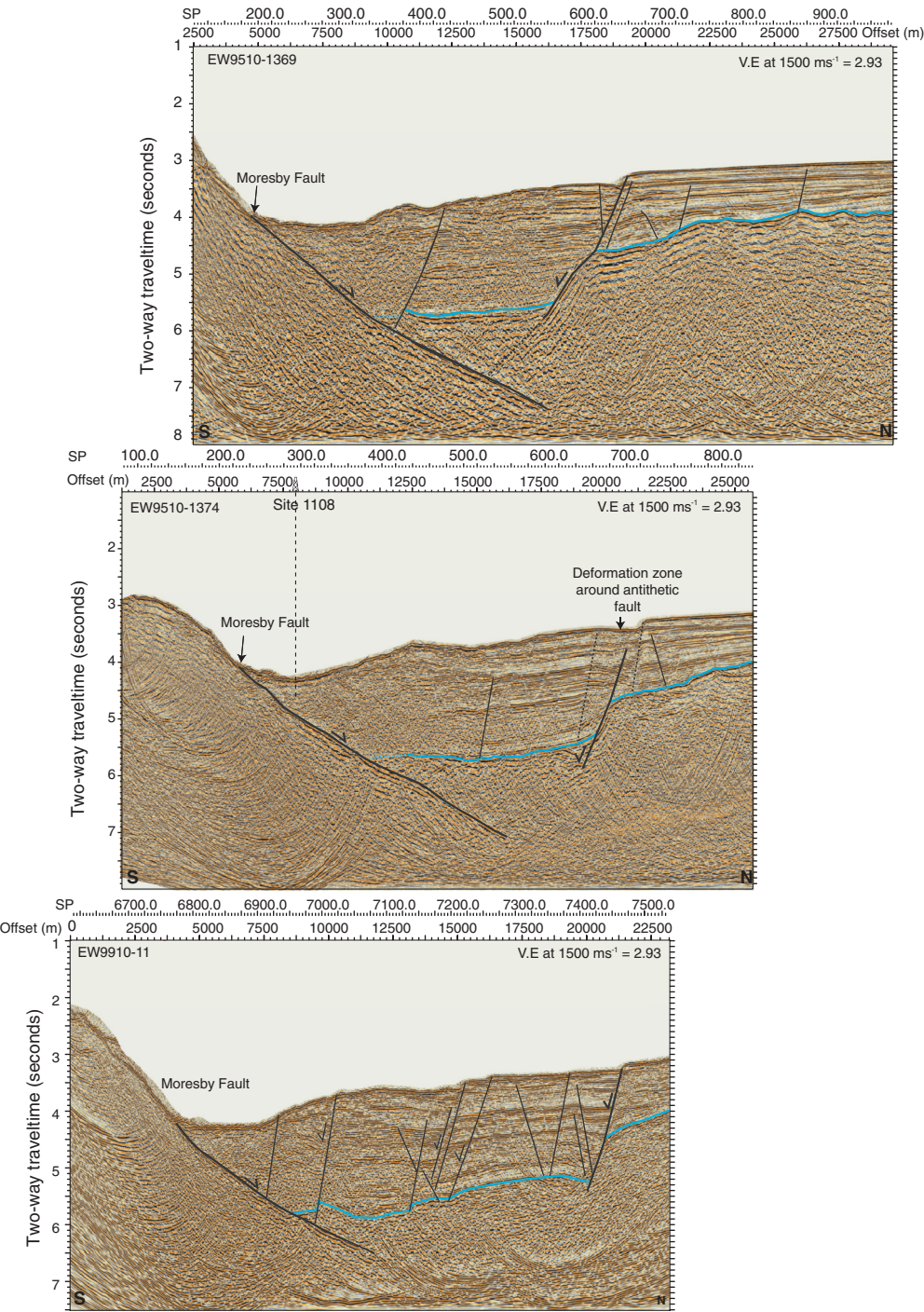


Fig. 6. MCS lines EW9510-1369 (top), EW9510-1374 (middle) and EW9910-11 (bottom) across the North Moresby Graben to the west of the rifting–spreading boundary (see Fig. 5 for locations). The MCS lines image the shallow-angle Moresby Fault and the high-angle, graben-bounding, antithetic fault system (on which the profiles are aligned). The blue line marks the base of the synrift sediments and the location of ODP Site 1108 is shown. For clarity, only some of the larger-offset normal faults are drawn. The NMG widens eastward; only the western line, EW9510-1369, images the Moresby Fault passing under the antithetic fault.

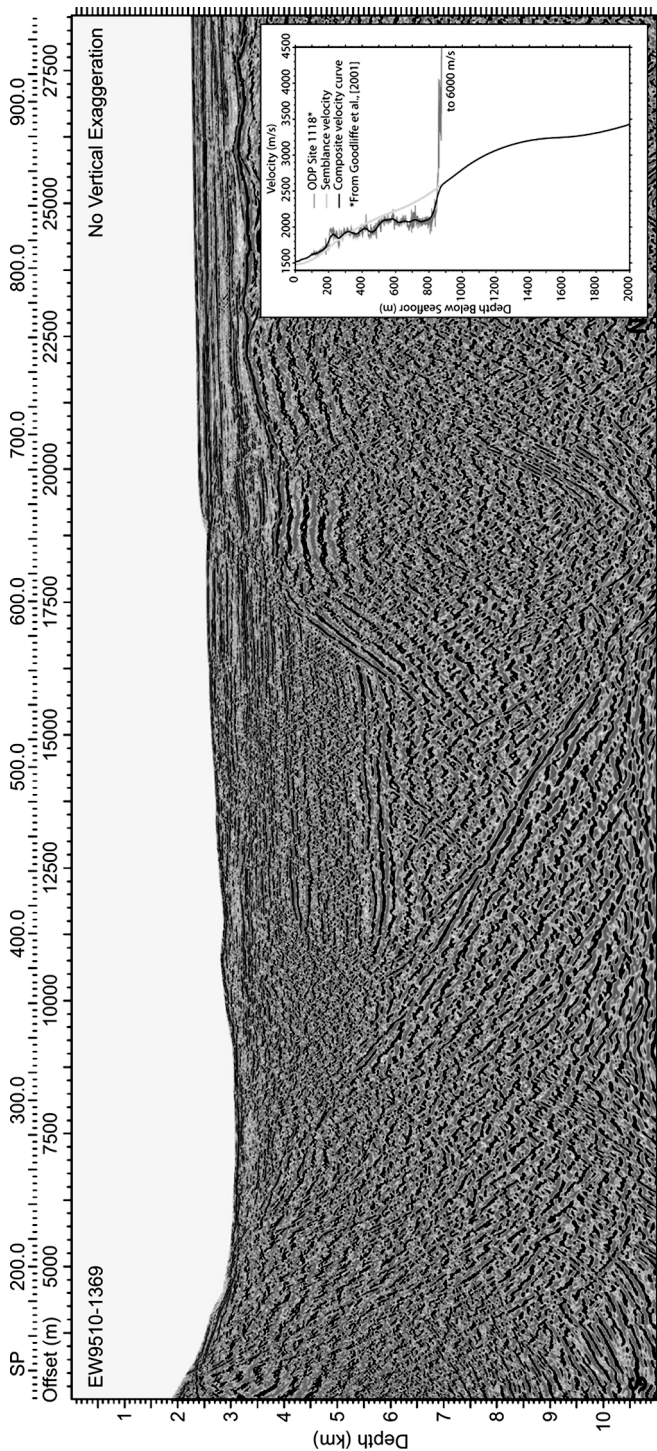


Fig. 7. MCS line EW9510-1369 (see Fig. 6a for time section) converted to depth using a velocity model for the synrift sediments derived from ODP results and semblance velocities. A velocity of 4.5 km s^{-1} is assigned to basement at the base of the synrift sediments. Below this a gradient is applied such that the velocity is 6 km s^{-1} at 8 s TWT. In this depth section the apparent dip of Moresby Fault is approximately 29° and of the antithetic fault is 54° .

line EW9910-11 (Fig. 6) where sediment onlap to the north is likely to have been the result of basement tilting to the south associated with early motion on the Moresby Fault.

The thick NMG sedimentary fill is cut by myriads of small-offset, north- and south-dipping, high-angle normal faults. For clarity, most of these faults are omitted on the interpretations in Figure 6, but the surface traces of the larger ones are shown in Figure 2. Sediment reflectors lose coherency as Moresby Fault is approached, usually above the point at which the basement reflector intersects the fault. This is probably the result of fault drag, extensional collapse, fluid flow and deposition of slump material.

Moresby Fault was first proposed to be active on the basis of teleseismic waveform inversions of normal faulting earthquakes in the vicinity of Moresby Seamount that had shallow-dipping nodal planes (Abers 1991). At least two of the focal mechanisms demonstrate normal faulting at dips of 25° – 35° , consistent with EW9203 MCS data (Abers *et al.* 1997). Evidence of water multiples in a water depth of 3 km indicated that the location of one of the largest earthquakes, on a plane dipping $31.6^{\circ} \pm 1.3^{\circ}$ N, is beneath the deepest part of the NMG and coincident with the Moresby Fault. An initial estimate of the dip of Moresby Fault made using MCS lines EW9510-1369 and 1374 was $27^{\circ} \pm 3^{\circ}$ towards 015° (Taylor *et al.* 1999).

Velocity data collected by logging and laboratory measurements during ODP Leg 180 allow a more accurate determination of the dip of Moresby Fault. Here we present a depth-converted version of MCS line EW9510-1369 using realistic velocities for the synrift sediments in NMG. This line has a number of desirable characteristics: (1) Moresby Fault and the south-dipping antithetic fault intersect at depth; (2) a distinct and continuous reflector is imaged along the length of the Moresby Fault; and (3) in two-way travel-time both faults have a distinct kink that is due to velocity pull-up where they descend below the keystone basement block that should be removed by depth conversion. Using the velocity model derived in Appendix 2, MCS line EW9510-1369 was converted to depth (Fig. 7). Moresby Fault has an average apparent dip of 29° N and that of the antithetic fault is 54° S. The average strike of the antithetic fault system is east–west (Figs 2 & 5) and hence its apparent dip on this line is true. In contrast, Moresby Fault has a sinuous fault trace (Figs 2 & 5), depth contours indicate a curvilinear form and varying strike down-dip (Taylor *et al.* 1999), and the direction of slip may have changed during its long history (Fig. 4). If an upper fault strike of 121.5° , determined from 11 200 m-spaced EW9910 MCS lines across the NE face (Fig. 5), is

used then the apparent dip equates to a true dip of 38° . However, if the long-term slip paralleled small circles about the main Euler pole of opening (i.e. $N7^{\circ}$ E, Fig. 2) then the Moresby Fault dip in this direction is 29.2° , and in the 015° direction used by Taylor *et al.* (1999) the dip is 29.9° . Fault dips of 30° – 33° are compatible with the slip plane of the large earthquake located beneath the NMG (Abers *et al.* 1997).

That the approximately 30° – 33° Moresby Fault is active has been the subject of much controversy. Theory states that if the coefficient of friction is that of typical rock (0.6–0.8) and σ_1 is near vertical then a normal fault will dip near 60° (Anderson 1951) or 45° for the case of finite slip (Forsyth 1992; Buck 1993). Lower-angle faults either may be explained by rotation of the stress field (Yin 1989; Parson & Thompson 1993), weakening through near lithostatic pore-fluid pressures within the fault zone (Byerlee 1990) or fault gouge materials with a lower coefficient of friction (Hill & Thatcher 1992). A primary goal of ODP Leg 180 was to penetrate Moresby Fault and investigate the mineralogy and stress state to determine why this fault is active. Although the active segment of Moresby Fault was not penetrated, 12 m of talc–chlorite–serpentine clayey fault gouge with coefficients of friction of 0.21–0.3 (Kopf *et al.* 2003) was recovered at Site 1117 along a segment of the fault exposed at the sea floor. Underneath, a poorly recovered approximately 50 m-thick section of mylonites (recording plastic deformation) and ultracataclastites (recording intermittent plastic and brittle deformation) was recovered from the quartz–gabbro basement. Analysis of samples from Site 1117 and other proximal sites has demonstrated the importance of frictional heating and hydrothermal fluid flow in the evolution of the Moresby Fault (Kopf *et al.* 2003). Channelling of fluids along this fault is evinced by the greater permeability of the fault gouge, especially in the direction parallel to the tectonic fabric, than the overlying sediments (Kemerer & Screatton 2001; Kopf 2001; Stover *et al.* 2001). The continuation of the fault gouge along the Moresby Fault beneath the NMG is indicated by inverse models of the MCS data that include a 4.3 km s^{-1} , approximately 33 m-thick layer along the fault with isolated pockets of a 1.7 km s^{-1} lower velocity material (Floyd *et al.* 2001).

Given the friction coefficients of the fault gouge and the presence of fluids in the fault zone, either of which alone would make motion on the Moresby Fault feasible (Abers 2001), strain accumulation on this fault is no longer an issue. The initiation of motion along an approximately 30° fault remains problematic, for although full flexural modelling of the evolution of rift systems shows the importance of rotation from initially high to low dips (Buck

1988), rotating the Moresby Fault to higher dips simultaneously rotates the large, contemporary, south-dipping normal fault within Moresby Seamount to low dips (Fig. 4) (Taylor & Huchon 2002). The abundant pre-existing structures in this once convergent regime (Davies & Jaques 1984) may play a role.

Morphology and structure at the point of continental breakup

Sea-floor spreading segment 1a occupies the southern half of the eastern NMG and obliquely truncates the SW end of Egum Graben. It is identified by a narrow band of high-amplitude positive magnetization and a coincident region of sea floor with high acoustic backscatter extending from 151°40'E to 151°52.25'E (Figs 2 & 5). At the eastern end of the segment, the 8 km-diameter Cheshire Seamount, bisected by an east–west cleft, rises more than 850 m above the surrounding sea floor to a depth of 1750 m. Recent lava flows extend down-slope to the east but are unlikely to be underlain by extensive intrusives given the reduced amplitude of the magnetization. Directly to the west of Cheshire Seamount, and at depths of 3050 and 3250 m, are two ring-shaped structures about 4 km in diameter that are in the centre of the Brunhes magnetization high (Figs 2 & 5). Around both constructs, high acoustic backscatter reveals lava flows that have pooled against the base and extended a short distance up the slopes of the surrounding fault-bounded ridges. Hyaloclastites were found in the sediments 5 km to the NW at ODP Site 1108 (Shipboard Scientific Party 1999b). They are high Na₂O (4.3%), high TiO₂ (2.8%) basaltic andesites (SiO₂ = 53%) that are light rare earth element (LREE)-enriched (chondrite-normalized La/Yb = 3.5) (Lackschewitz *et al.* 2001). Similar rocks have been dredged from Cheshire and Franklin seamounts; the former also includes dacites (Binns & Whitford 1987; Dril *et al.* 1997). Other rocks from the west end of segment 1a are normal mid-ocean ridge basalts (N-MORBs) with indications of low degrees of melting (at MgO = 8%, Na₂O = 3.1%, FeO = 8.1% and TiO₂ = 1.14%) (Binns & Whitford 1987). Thus, the mantle sources of segment 1 lavas include N-MORB, arc-like and enriched components (Dril *et al.* 1997).

The extent of the intrusive body that fed the segment 1a lavas can be derived from the magnetization anomaly. Assuming that the intrusion is broadly tabular, the edges of the body will be coincident with peaks in the horizontal derivative of the anomaly (Goodliffe 1993). A series of north–south profiles across the magnetization anomaly were extracted and the horizontal derivative calculated and maxima plotted. Five profiles centred on two locations (A1–A2, B1–B2; Fig. 5b) were stacked to remove

noise before the maximum horizontal derivative was calculated. The horizontal derivative for the western part of segment 1a displays distinct peaks that lie on either side of a body that is between 4.1 and 5.7 km wide. Closer to the Cheshire Seamount the calculated width of the body increases to 7 km, which suggests that segment 1a propagated to the west.

Using the current Euler pole and an intrusive body width of 5 km we calculate that sea-floor spreading began on the western part of spreading segment 1a by 0.11 Ma. This is considered an upper limit as some of that width may originate from overprinting rather than strain accommodation. For example, if spreading initiation coincided with the 80 ka reorientation of the entire Woodlark Basin spreading centre (Goodliffe *et al.* 1997) then only 3.6 km of the intrusion would represent strain accommodation.

MCS line EW9910–13 crosses the western end of segment 1a, providing a snapshot of the development of this young accretionary plate boundary (Fig. 8). On this MCS line, the Moresby Fault and the antithetic fault that together bound the NMG are readily identified, as are the thick synrift sediments above the keystone basement block. Between SP8300 and SP8500, however, there is a region of incoherent reflectivity below 5 s two-way time (TWT) against which the Moresby Fault, the synrift sediments and the keystone basement reflectors terminate (Fig. 8). We interpret this incoherent region to be the expression of the Brunhes axial intrusion. It correlates closely with, but is about 800 m north of, the limits of the intrusive body determined from the magnetization anomaly (Figs 5 & 8). Those limits exactly match the position of three sea-floor ridges that, with the bathymetry data, we interpret as volcanic piles above a ring dyke (northern two ridges) plus another dyke to the south, fed from the intrusion below (Fig. 8).

Above Moresby Fault at the southern end of MCS line EW9910–13 there are few coherent reflectors, perhaps owing to deposition of sediments in the form of slumps originating on Moresby Seamount (probably similar to talus found at ODP Leg 180 Site 1108) (Shipboard Scientific Party 1999a) (Fig. 8). To the north of the intrusive body the synrift sedimentary sequence is offset at SP7925 by a south-dipping normal fault. This fault is imaged on MCS line EW9910–11 to the west and continues to the east as the major northern bounding fault of Egum Graben. The synrift sequence is interpreted to be continuous with and have a depositional history similar to that sampled to the west during ODP Leg 180 (Shipboard Scientific Party 1999a).

The evidence presented above indicates that a major intrusive body underlies the centre of

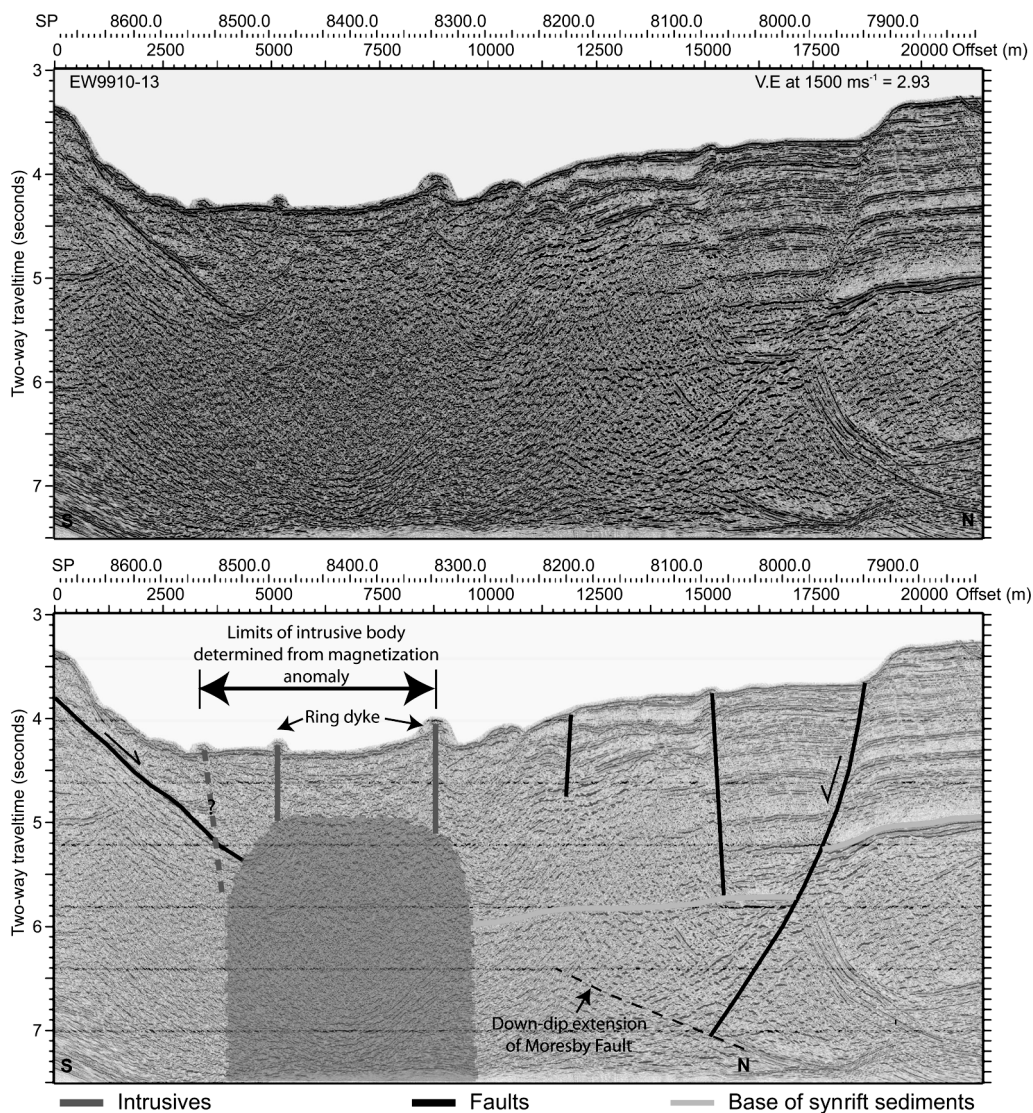


Fig. 8. Uninterpreted (top) and interpreted version (bottom) of MCS line EW9910-13 (see Fig. 5 for the location). The currently forming continent–ocean boundary involves igneous intrusions penetrating the shallow-angle normal fault of an asymmetric rift graben. Moresby Fault is cut by an approximately 5 km-wide intrusive body that is overlain by subhorizontal sediments that have been cross-cut by a ring dyke.

segment 1a. With a width of 5 km it is assumed that it is today accommodating extension in a similar fashion to a mature spreading centre, although the intrusion has only broken the surface in the form of localized dykes. If the intrusive body is similar to a mature spreading centre it may take the form of a sheeted dyke sequence that is analogous to those observed at ophiolites and inferred to exist below modern spreading centres. Although there is no evidence of an active magma chamber, the sheeted dyke sequence may be underlain by massive

gabbros. Although there are extensive lava flows at the surface, the MCS data suggest that they overlie an approximately 500 m-thick (assuming a velocity of 1800 m s^{-1}) extension of the synrift sedimentary sequence seen to the north. As the intrusive body does not reach the surface, extension above it must be transferred into the overlying synrift sediments. The well-stratified reflectors show little evidence of brittle faulting, implying either a dominance of soft-sediment deformation or the offset of extension to either side of the intrusion. Much of the sediment

in this bathymetric low may have been deposited rapidly and recently.

Discussion

Breakup model

We present a simplified model for the formation of the current rift-spreading boundary in the Woodlark Basin (Fig. 9). The initial condition, prior to continental extension, would resemble the central

Papuan Peninsula today, with an orogenic mountain chain of 1–3 km elevation, crust about 40 km thick, arc volcanoes spaced along the northern coastal plain and a broad shallow-marine forearc basin south of the Trobriand Trench (Taylor *et al.* 1999). At 8.4 Ma (Taylor & Huchon 2002) this inherently asymmetric region started to undergo north–south extension, lowering the topographic profile (Fig. 9a) and exhuming the metamorphic core complexes of the D'Entrecasteaux Islands (Fig. 1) to the west. Successive generations of

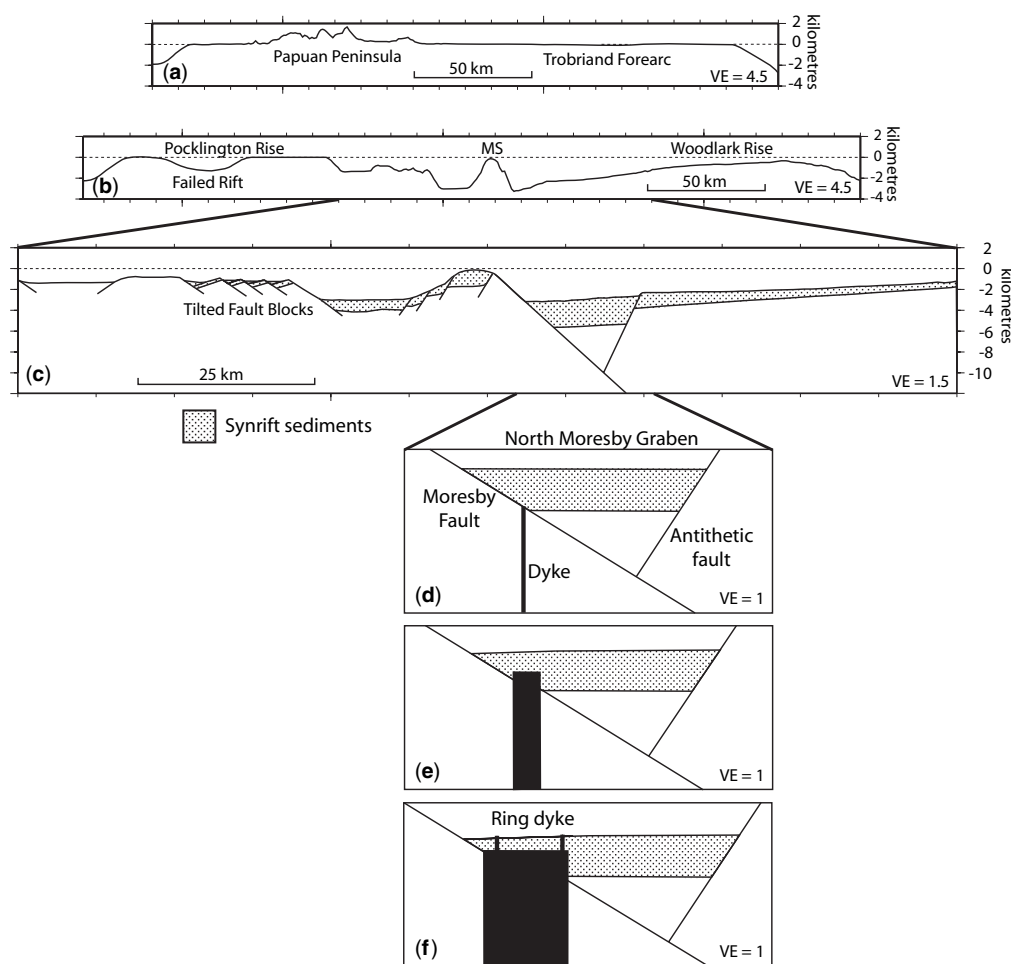


Fig. 9. The evolution of continental rifting through sea-floor spreading in the western Woodlark Basin. The Papuan Peninsula (topography based on a north–south profile across the Papuan Peninsula at 148.8°E: (a) has undergone broadly north–south extension, subsiding to current depths at the longitude of Moresby Seamount (b). Southern graben represent former loci of extension that are now failed rifts. The southern margin is characterized by large fault blocks rotated by high-angle faults. The northern margin has subsided with little evidence of large-offset faults (c). At about 1.2 Ma extension focused on South and North Moresby Graben. More recently, extension has focused at the NMG and, ultimately, to Moresby Fault. A single dyke was intruded into the fault and the overlying sediments causing motion on the up-dip segment of Moresby Fault to cease (d). Progressively growing in width (e), the intrusion is today about 5 km wide but has only reached the surface in the form of localized dykes (f) and flows.

normal faults thinned the palaeo-Papuan Peninsula, resulting in subsidence below sea level to form the Pocklington Rise. To the north, the palaeo-Trobriand forearc (now Woodlark Rise) has tilted to the south and subsided by up to 3 km, reversing the northward-directed Miocene drainage pattern (Fig. 9b). One model for the growth of the metamorphic core complexes in the D'Entrecasteaux Islands (Fig. 1) suggests that, with the aid of fluids liberated from the subducted Solomon Sea Plate, the lower crust of the northern margin is more capable of ductile flow than the margin to the south (Martinez *et al.* 2001). The paucity of upper crust normal faults in the northern margin implies that crustal thinning must be primarily achieved through means other than brittle faulting. The bulk of the thinning may take place through the removal of the lower crust (Driscoll & Karner 1998; Buck 2004; Kusznir *et al.* 2004). In contrast, brittle failure of the upper crust is more prevalent to the south. The margin asymmetry need not reflect lithospheric-scale simple shear, but instead is a function of the prerift geology, morphology and rheology.

The broadly distributed extension of the palaeo-Papuan Peninsula included locally focused extension that produced large graben in the southern margin that are now failed rifts (Figs 1 & 9b). Tilted fault blocks overlain by rotated sediment packages in broadly equal-sized graben (SCS line MW9304-30, Fig. 3) show that extension only focused on the axial graben very recently, even though results from ODP Leg 180 (Taylor & Huchon 2002) show that the latter formed early in the rift history (Fig. 4). A broad heat flow anomaly, that is centred in the northern part of the SMG but also covers the NMG, reflects maximum crustal thinning across this axial system (Goodliffe *et al.* 2000). Today, as evinced by seismicity (Fig. 1), extension has largely ceased on the southern faults, and is concentrated on the shallow-angle north-dipping Moresby Fault (Taylor & Huchon 2002). MCS lines EW9510-1366 (Fig. 4), EW9510-1369, EW9510-1374 and EW9910-11 (Figs 6 & 7) reveal the final phase of current rifting prior to breakup. MCS lines EW9510-1366 and EW9910-11 show synrift sediments in the NMG dipping to the south owing to slip on the Moresby Fault, whereas the keystone basement block and overlying sediments in the NMG are less rotated on the intervening lines.

The stage where extension is primarily accommodated by accretion is represented by MCS line EW9910-13 (Fig. 8). To reach this stage, we envision that Moresby Fault was cut by a dyke swarm that gradually grew in width (Fig. 9d, e). When the first dykes cut through the fault and were emplaced into the overlying synrift sediments, motion on the up-dip portion of Moresby Fault ceased and/or shifted upwards into the sediments. A transfer fault/zone may have formed to connect accretion on spreading segment 1a to normal

faulting on Moresby Fault in the NMG. The NE flank of Moresby Seamount may be such a transfer fault; its NW–SE trend is consistent with small circles about the current Euler Pole (12°S , 144°E , $2.437^{\circ}\text{Ma}^{-1}$, Fig. 2b). If the width of the intrusive body represented only strain accommodation through accretion, the Moresby Fault would exit the intrusive body at the same height on the northern side as it entered on the southern side. Although there is some ambiguity in the location of the continuation of Moresby Fault to the north of the intrusion, this is clearly not the case (Fig. 8). Although we maintain that the majority of the intrusive body represents strain accommodation through accretion, which formed the associated magnetic anomaly (Fig. 5), a portion of the width may represent overprinting of the original country rock, perhaps through a process similar to stoping. Alternatively, the intrusion may not be continuous, but instead have intervening unresolved slices of country rock between dykes. For the purposes of our schematic model, we have arbitrarily assigned approximately 25% of the intrusion width to these processes (Fig. 9e, f). As the intrusion grew, sediments and possibly sills were added to the centre of the graben. A ring dyke was intruded and reached the sea floor (Fig. 9e) leading to eruptions that produced the extensive lava flows observed (Fig. 2c). Continued extension, vertical propagation of the intrusive body and horizontal growth of the ring dykes will soon construct an edifice at the sea floor on the western (as already exists on the eastern) part of spreading segment 1a.

As shown here, the currently forming continent–ocean boundary involves igneous intrusions penetrating the principal (and shallow-angle) normal fault of an asymmetric rift graben. Further east in the Woodlark Basin, breakup along spreading segment 2 occurred within symmetric graben bounded by high-angle faults (Taylor *et al.* 1999). But in each case the magnetic, bathymetric and seismic data reveal that the boundary between continental rifting and sea-floor spreading in the Woodlark Basin is locally very sharp—notwithstanding that regionally the processes overlap in time. This is in stark contrast to breakup of the Newfoundland and Iberia conjugate margins associated with ultra-slow spreading in the early north Atlantic where transition zones up to 150 km wide separate clearly defined oceanic and continental crust (Tucholke *et al.* 2004).

Margin stratigraphy and unconformities

MCS imaging calibrated by ODP drilling results allow us to observe the stratigraphic details of continental breakup in the western Woodlark Basin. How do these images and results compare to stratigraphic models of continental margin development?

Falvey (1974) proposed a model to explain typical patterns of margin uplift and subsidence as a function of the thermal, erosional and metamorphic processes attendant on splitting a continent by emplacing and then cooling asthenosphere. He observed that two angular unconformities commonly separate pre-, syn- and post-rift sediments. He proposed that 'the breakup unconformity is caused by erosion during the final uplift pulse associated with prebreakup upwelling in the mantle'. He recognized that 'this unconformity is more localized than the rift-onset unconformity, being difficult to define in troughs' and that 'continental breakup and the onset of sea-floor spreading is probably not instantaneous'. The existence of two unconformities bounding synrift strata has found widespread application in rifted margin studies (Driscoll *et al.* 1995) even though, in the particular example Falvey used to illustrate the general model (the Otway Basin of South Australia), the proposed rift-onset unconformity actually correlates with the time of breakup and initiation of marine sedimentation, and the real rift-onset unconformity occurred much earlier (Hegarty *et al.* 1988).

In the case of the western Woodlark Basin, there is a regional rift-onset unconformity. It truncated mafic igneous basement at ODP sites 1109 and 1118, produced a 4 Ma hiatus in the shallow-marine forearc basin sediments at ODP Site 1115 (Figs 3 & 4), and generated the late Miocene 'unconformity A' in the Nubiam-1 well and region north of the D'Entrecasteaux Islands (Francis *et al.* 1987; Taylor & Huchon 2002). The drilling and MCS studies show the reversal of the northward Miocene forearc drainage as the Papuan continent was stretched and thinned, and the northern margin (Woodlark Rise) progressively subsided southward toward the active rifts, from paralic to shelf to the present bathyal water depths (Fig. 4) (Taylor & Huchon 2002). Concurrently, the former landmass to the south was faulted and thinned to produce the graben and reefs of the Pocklington Rise (Fig. 1).

Early synrift sediments were subsequently tilted on rotating fault blocks, but later sediments of the southern margin were ponded in half-graben as the early rifts in the south were abandoned, even as deformation continued further north focused on the axial rifts (Figs 3 & 9). Therefore 'synrift' sediments (those deposited between rift onset and breakup) can be characterized by rotated sections (with parallel or diverging reflectors) as well as by ponded sections that onlap bounding fault blocks where local extension has ceased. Splayed reflectors representing sediment deposition on rotating fault blocks do not characterize the full synrift section, but only those times and places where extension is localized. Also, in the common case where breakup propagates, the time represented by the synrift sediments and the timing of any breakup unconformity will vary

along strike: by 6 Ma in the Woodlark Basin. Furthermore, as Falvey (1974) noted, sedimentation in the deep water parts of margins is likely to be conformable across the time of breakup. In the Woodlark Basin, the progression from rifting to breakup and then initial spreading is characterized by a decrease in sedimentation as the margins are progressively thinned, subside below sea level, are eroded less and trap sediments in proximal basins (Figs 3 & 4). The post-rift sedimentation near the continent-ocean boundary is hemipelagic and drapes breakup topography. A distinct breakup unconformity is not observed.

Conclusions

- The continental margins asymmetry at the point of current breakup – high-angle normal faults bounding tilted crustal blocks in the south in contrast to a wide tilted margin in the north with only small-offset normal faults – cannot be explained by the observed shallow-angle north-dipping detachment between them. Rather, the margins asymmetry is a function of prerift geology, morphology and rheology. Removal of the lower crust through ductile flow is the preferred mode of crustal thinning of the Woodlark Rise, the palaeo-Trobriand forearc. In contrast, on the Pocklington Rise to the south, where the high mountains of the Papuan Peninsula existed 8.4 Ma ago, brittle faulting of the upper crust is more prevalent.
- Initially, major upper crust normal faulting was distributed from the southern margin to what is now South and North Moresby Graben. Sequentially, the locus of faulting narrowed northwards, leaving large graben (foci of former extension that are now failed rifts) and tilted fault blocks in the southern margin.
- At about 1.2 Ma, extension focused at NMG and SMG where rapid subsidence was accompanied by rapid uplift of the intervening Moresby Seamount. This was a precursor to the initiation of sea-floor spreading at approximately 0.8 Ma on segment 1. Extension ceased on the large faults bounding tilted blocks directly to the south.
- Sea-floor spreading on segment 1a initiated at approximately 0.08 Ma in the along-strike continuation of the NMG and more specifically through the north-dipping Moresby Fault. Ahead of the spreading tip, extension has focused on the active Moresby Fault (Abers *et al.* 1997; Taylor & Huchon 2002). Although motion on this normal fault was once controversial owing to its low dip, fault gouge with a low coefficient of friction and evidence of high pore fluid pressures make motion on this fault unremarkable (Taylor & Huchon 2002). Depth conversions of MCS line EW9510-1369 based

on ODP results and semblance velocities (Goodliffe *et al.* 1999b) indicate that the true fault dip is $\geq 29^\circ$ and compatible with an earthquake slip plane of 30° – 33° (Abers *et al.* 1997).

- The normal faults bounding the northern side of spreading segment 1, co-polar with great circles about the pre-0.52 Ma Euler pole (Fig. 2), are crossed by the Egum Graben. The faults bounding the northern side of Egum Graben form the antithetic faults in the NMG and may be continuous with a major transform fault system that separates the Solomon Sea and Woodlark lithosphere to the east. If so, this would make the current rifting–spreading transition a triple junction between the Trobriand, Woodlark and Australian plates. Alternatively, the fact that the bounding faults of Egum Graben are co-polar with great circles about the present-day Euler pole suggests the possibility that this graben represents a new rifting trend, perhaps reactivating pre-existing weaknesses.
- MCS line EW9910-13 records an early evolutionary step in the initiation of a spreading centre. An intrusion that cuts an along-strike continuation of Moresby Fault is connected to the surface via a ring dyke through which lavas have erupted onto the sea floor. Thus, the currently forming continent–ocean boundary involves igneous intrusions penetrating the principal (shallow-angle) normal fault of an asymmetric rift graben. The magnetic, bathymetric and seismic data reveal that the boundary between continental rifting and sea-floor spreading in the Woodlark Basin is spatially very sharp – notwithstanding that temporally the processes overlap.
- Synrift sediments (those deposited between rift onset and breakup) can be characterized by rotated and/or diverging sections, as well as by ponded sections that onlap bounding fault blocks where locally extension has ceased. A breakup unconformity is not generally observed at the deep-water boundary between rifted continental and accreted oceanic crust, but rather there is a decrease in sedimentation as the margins are progressively thinned, subside below sea level, are eroded less and trap sediments in proximal basins.

Appendix 1: Data and processing

The data presented herein come primarily from three marine geophysical surveys conducted between 1993 and 1999 (Goodliffe *et al.* 1999b) and include single-channel seismic (SCS) data, multichannel seismic (MCS) data, swath bathymetry, acoustic imagery and magnetic data. Generic mapping tools (GMT) (Wessel & Smith 1995) was used comprehensively throughout this study both for display and analysis.

Bathymetry and acoustic imagery data

HAWAII-MR1 swath bathymetry and sidescan data from a 1993 R/V *Moana Wave* survey cover the western Woodlark Basin from $150^\circ 50'E$ to $154^\circ 45'E$. Hydrosweep multibeam bathymetry data from R/V *Maurice Ewing* cruises in 1995 and 1999 provided additional detail in the region of Moresby Seamount. Beyond the coverage of the swath datasets, wide-beam profiler data from the National Geophysical Data Center (NGDC) database and digitized soundings from Australian Navy charts were incorporated. Land topographic data are from the Shuttle Radar Topography Mission (SRTM) 90 m grid cell dataset. Final bathymetry grids were produced with a 0.001° grid size.

Magnetic data

Magnetic data were collected along north–south 5 nm-spaced tracks during the regional 1993 R/V *Moana Wave* survey and during R/V *Maurice Ewing* surveys in 1995, 1999 and 2000 in the vicinity of Moresby Seamount. As the magnetic anomalies formed along an east–west spreading centre are highly skewed, the shipboard magnetic and bathymetry data were gridded at 0.01° and used to derive a solution for the intensity of magnetization of the sea floor following the technique of Macdonald *et al.* (1980). The magnetization inversion assumes a 1 km-thick source layer conforming to the sea floor. As this solution produced a reversal pattern in qualitative agreement with that from the total field anomalies, no annihilator (Parker & Huestis 1974) was added to the solution. Areas shallower than 1.2 km were masked to satisfy the requirement that the magnetic source does not come within one grid cell of the observation surface (Parker & Huestis 1974). Although the resulting magnetization map can be interpreted directly using the geomagnetic polarity timescale (Cande & Kent 1995), the oceanic crust in the region of interest is mostly formed during the Brunhes chron.

Marine seismic data

The marine seismic data used in this study comes from three surveys. In 1993 six-channel seismic (SCS) data were collected during an R/V *Moana Wave* survey in the Western Woodlark Basin using a 150 m-streamer and a 120 cubic inch (inch³) airgun source. Shots spaced at 37.5 m yielded twofold 12.5 m-spaced CMPs. A 1995 R/V *Maurice Ewing* survey included a site survey for ODP Leg 180 (Taylor *et al.* 1996; Goodliffe *et al.* 1999b). To better image the Moresby Fault and document the subsidence of the northern margin of the Woodlark Basin, a 196-channel 4.9 km-streamer was deployed. The first line, designed for deepest penetration, was shot using a tuned 8460 inch³ 20 airgun array. Shots every 20 s yielded a shot spacing of approximately 50 m and 48-fold 12.5 m-spaced CMPs, a record length of

16.384 s and a sample interval of 4 ms. Additional lines were shot using 10 airguns alternating in the same 20 airgun array. Shots every 12 s yielded a shot spacing of approximately 25 m and 60–75-fold 12.5 m-spaced CMPs with a record length of 9.116 s and a sample interval of 4 ms. In 1999 the R/V *Maurice Ewing* shot 13 dip lines oriented NNE–SSW, perpendicular to the strike of the fault (Taylor *et al.* 1999) and one strike line. Eleven of the dip lines, spaced 200 m apart, crossed Moresby Graben in the vicinity of ODP Site 1108. The two remaining dip lines crossed the western tip of the spreading centre to the east. Using a 48-channel 1.2 km-streamer, a 1395 inch³ tuned six airgun array was fired every 25 m, giving a 24-fold CMP interval of 12.5 m. The record length was 8192 ms and sample interval was 2 ms.

The marine seismic data were processed using ProMAX. Data processing included trace editing, filtering, prestack and post-stack deconvolution, velocity analysis, inside and outside mutes, normal moveout corrections, stack and migration. The multichannel seismic (MCS) data presented herein have also undergone partial prestack migration.

Appendix 2: Velocity model and depth conversion

Four sites were drilled and logged in the synrift sedimentary section. At Site 1108, sampled to a depth of 480 m, density and porosity measurements showed that 385 m of sediments have been removed from the top of the section through faulting and/or erosion (Shipboard Scientific Party 1999b). The termination of reflectors at the sea floor south of SP420 on MCS line EW9510-1374 and south of SP400 on MCS line EW9510-1369 (Fig. 6) confirms that this is a reasonable interpretation and any synrift sediment velocity model must take this into account. To the north of the NMG logging and laboratory measurements of velocity were obtained at sites 1109, 1115 and 1118. Depth v. velocity curves were derived for each of the three sites, refined using vertical seismic profile checkshots, and verified through the generation of synthetic seismograms (Goodliffe *et al.* 2001). Site 1118, on the margin just north of the NMG (Figs 2 & 4), sampled 860 m of synrift sediments. Numerous MCS lines (Goodliffe *et al.* 1999b) show that the sediments sampled on the northern margin can be correlated with those in NMG close to MCS lines EW9510-1369 and EW9510-1374. Here, the offset across the northern bounding fault is small, enabling the velocity–depth curve for Site 1118 to be used in the graben. As we only have synrift sediment velocity information to a depth of 801 m, the best estimate of interval velocity for deeper synrift sediments in the graben comes from semblance analysis on line EW9510-1369 where the strata are continuous and subhorizontal. The composite interval velocity curve predicts a velocity of approximately 3.5 km s⁻¹ at about 2.5 km below the sea floor (the approximate sediment thickness in the graben). Shipboard laboratory measurements of the basement

diabase consistently gave velocities of approximately 6 km s⁻¹ (Shipboard Scientific Party 1999a). As fracture porosity of this lithology is often high, we assigned a velocity of 4.5 km s⁻¹ to the basement at the base of the synrift sediments and applied a gradient such that the velocity would be 6 km s⁻¹ at 8 s TWT directly below the centre of the graben.

Extrapolating our one-dimensional velocity model to two dimensions in most places entailed hanging the sediment and basement velocity functions from the sea floor and basement, respectively. However, south of SP400, where reflectors terminate at the sea floor, the composite velocity curve was suspended at 3.54 s TWT, effectively increasing the sediment velocity at the water–sediment interface and closely matching the velocities of the sediments recorded at Site 1108. The water column was assigned an average velocity of 1494 m s⁻¹.

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