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CHAPTER 26—Microcontinent formation around Australia

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Microcontinents are common in the accreted continental geological record, but relatively rare in modern settings. Many of today's microcontinents are found in the Tasman Sea and Indian Ocean. These include the East Tasman Rise, the Gilbert Seamount Complex, the Seychelles, Elan Bank (Kerguelen Plateau), and possibly fragments of the Lord Howe Rise and Norfolk Ridge, and the Wallaby Plateau. We review their history of formation, and propose that the mechanisms that led to their isolation were mostly plume-related. Tasman Sea continental fragments formed by ridge jumps onto adjacent continental margins after sea-floor spreading in the southern Tasman Sea commenced. The East Tasman Plateau was separated from the Lord Howe Rise at about chron 34 (83 Ma) and the Gilbert Seamount Complex rifted off the South Tasman Rise at roughly 77 Ma, by ridge jumps in opposing directions. Evidence for thermal anomalies under the central Lord Howe Rise, Ross Sea and possible eastern Australian margin explain ridge jumps that led to the isolation of the East Tasman Plateau, Gilbert Seamount and possibly the northern Lord Howe Rise and Dampier Ridge. In the central Indian Ocean, spectacular exposures of granite make the Seychelles a type example of a microcontinent. As in the Tasman Sea, ridge–plume interactions have been responsible for separating a thinned continental sliver from a large continent (India). Elan Bank, as part of the Kerguelen Plateau, represents another example of a continental fragment in the Indian Ocean. Newly identified M-sequence anomalies in the Enderby Basin, off Antarctica, suggest that this microcontinent was detached from India no earlier than 124 Ma when a northward ridge jump towards the Kerguelen plume may have isolated Elan Bank. This interpretation implies that a mantle thermal anomaly due to the incipient Kerguelen plume pre-dated the Rajmahal Traps, emplaced at about 118 Ma, by ~6 million years. An early Kerguelen hot-spot position north of Elan Bank, with subsequent southward migration, is also supported by palaeomagnetic data and mantle convection models. The Wallaby Plateau off Western Australia, located between the Perth and Cuvier Abyssal Plains, may also include slivers of continental crust, but unequivocal evidence is not available, as it has never been drilled. Here it is suggested that most of these microcontinents formed by re-rifting of a young continental margin in the vicinity of a mantle plume stem. The weak inner flank of a rifted margin weakens further when passing over a mantle plume, causing a nearby spreading ridge to jump onto this zone of weakness. This process isolates a passive margin segment, and leaves a narrow passive margin behind. After the onset of subduction and ocean basin destruction, such microcontinents may be accreted again to an active plate margin.

KEY WORDS: hot spots, microcontinent, rifting, sea-floor spreading.

INTRODUCTION

Many of today's microcontinents are found in the proximity of Australia, in the Tasman Sea and Indian Ocean (Figure 1). In the geological past microcontinents contributed to continental growth as accreted terranes, especially in eastern Australia (Crawford *et al.* 2002; Veevers 2000).

Several mechanisms have been proposed to explain the formation of microcontinents. Vink *et al.* (1984) argued that if rifting occurs close to the boundary between continental and oceanic crust, the rift would preferentially form on continental crust because of its greater rheologic weakness. Steckler and ten Brink (1986) suggested that a rheologic strength minimum forms landward of the hinge zone of a rifted margin, a situation that favours the creation of thin continental slivers if renewed rifting occurs. Whether and when a lateral migration of a rift axis to a yield-strength minimum adjacent to the rift would take place depends on the lithospheric temperature, strain rate, and the reactivation time of extension (Negredo *et al.* 1995). Negredo et al's (1995) model accounts for multiphase rifting, but not for the jump or propagation of a spreading ridge into a continental margin after the onset of sea-floor spreading. An existing mid-ocean ridge should always yield to extension before re-rifting of continental crust would take place, as the yield strength of a mid-ocean ridge is extremely low (Bodine *et al.* 1981). Extension of a continental margin nearly always ceases after the onset of sea-floor spreading, causing a breakup unconformity (Braun & Beaumont 1989), as post-breakup divergence between two plates is accommodated by sea-floor spreading. Müller *et al.* (2001) identified four microcontinents which are exceptions to this rule: the Jan Mayen microcontinent in the Norwegian–Greenland Sea, the Seychelles

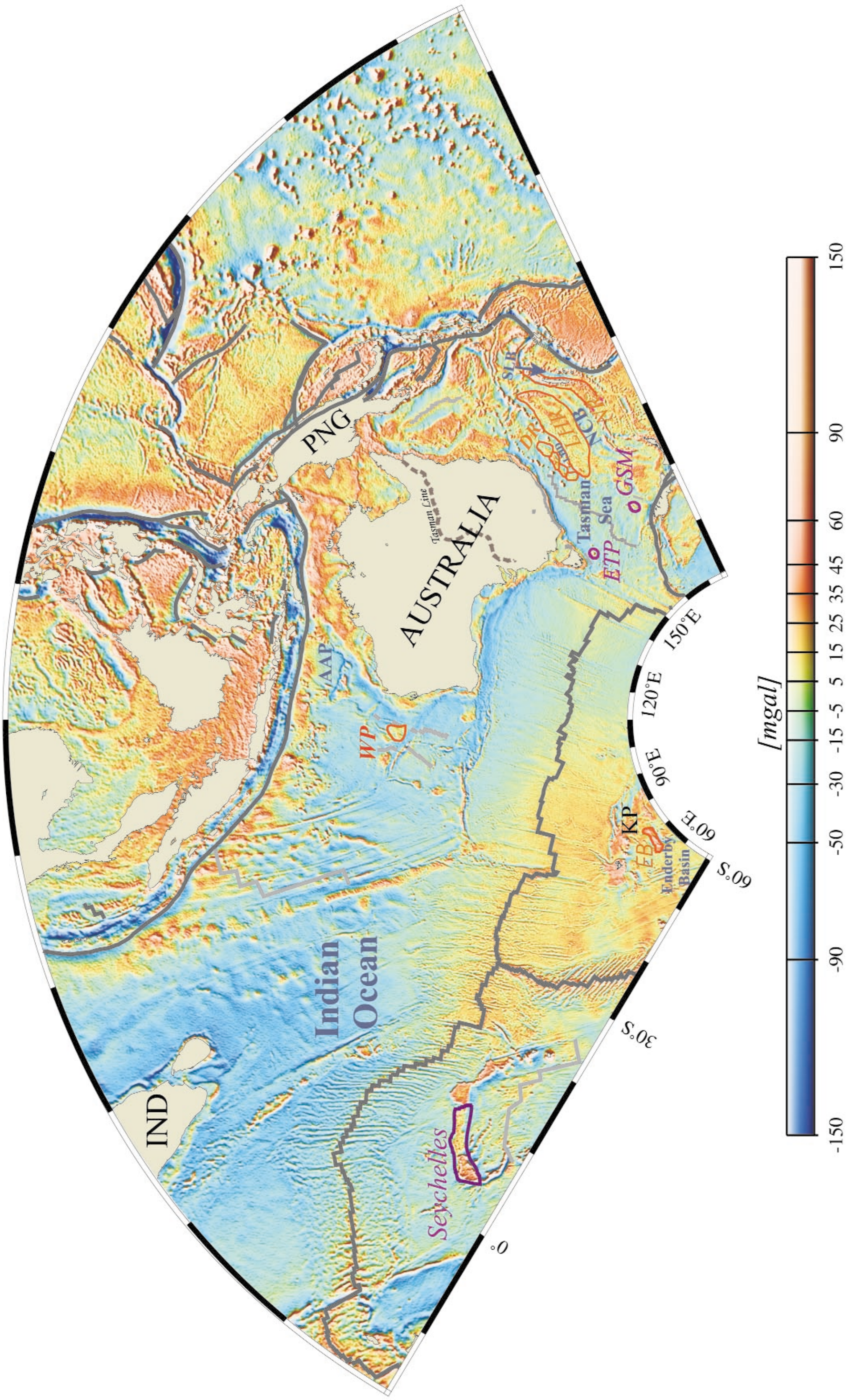

Figure 1 Gravity anomaly derived from satellite altimetry (Sandwell & Smith 1997) of oceanic areas around Australia. Microcontinents formed by hot-spot–ridge interaction mechanism are outlined in magenta or in red (for unconfirmed continental origin or mechanism: see text). ETP, East Tasman Plateau; GSM, Gilbert Seamount; LHR, Lord Howe Rise; DR, Dampier Ridge; NCB, New Caledonia Basin; SLB, South Loyalty Basin; LHD, Lord Howe and Dampier Basins; NZ, New Zealand; EB, Elan Bank; KP, Kerguelen Plateau; WP, Wallaby Plateau; AAP, Argo Abyssal Plain. Present-day plate boundaries are represented by thick, grey lines; extinct ridges are drawn in light grey.

in the Indian Ocean, and the East Tasman Plateau and the Gilbert Seamount Complex in the Tasman Sea (Figure 1). In these places, an existing spreading ridge propagated or jumped to a position within the continental margin, severing a small segment of stretched crust from a large continent. They showed that both microcontinent formation and nearby asymmetries in oceanic crustal accretion were triggered by plumes.

Apart from rifting, strike-slip and subduction processes can also initiate microcontinent formation. Margin-parallel shear may generate thin slivers that may move past their parent continent for thousands of kilometres (Sengor & Dewey 1991) as in the case of southeast Asia (Metcalfe 1988) and northern Australia (Pigram & Davies 1987). In extensional arcs, diverse buoyant slivers may be created by pulling out bits of pre-existing continental margin in the form of migratory ensialic arcs (Sengor & Dewey 1991). Transition from subduction to rifting associated with subduction of an active spreading centre, can also trigger continental block isolation, such as Baja California (Atwater 1970). Here we examine microcontinent formation around Australia and investigate the timing and mechanisms of their formation.

MODEL FOR PLUME-RELATED MICROCONTINENT FORMATION

A sequence of six events may lead to the formation of plume-related microcontinents: (i) emplacement of a large igneous province above a mantle plume head; (ii) rifting and sea-floor spreading between two large plates; (iii) the propagation or jump of the active ridge from the spreading axis to the landward edge of a young continental margin (less than 25 million years), leading to renewed rifting; (iv) minor volcanism; (v) sea-floor spreading and separation of a microcontinent; and (vi) a prolonged period of asymmetric sea-floor spreading (Müller *et al.* 2001) (Figure 2). The lack of substantial volcanism at and/or after the time of microcontinent formation may simply be a consequence of an existing plume stem, rather than a plume head, triggering the re-rifting of a young continental margin. Hot-spot activity may also be episodic (O'Connor *et al.* 2000) or a plume head may break into diapirs due to tilting in the upper mantle (Cox 1999). Therefore, according to the geometry and behaviour of the plume, microcontinent formation is not always contemporaneous with, or accompanied by, substantial volcanism. Plume-related microcontinent formation by a large ridge jump could be preceded by the formation of a large igneous province and/or volcanic-margin formation, reflecting plume-head emplacement. The association of large igneous provinces and microcontinents can be used to identify similar events in the geological record.

Re-rifting along a margin's inner edge can only occur if its yield strength is weakened such that it is at least as weak as a mid-ocean ridge. Without severe reheating of the margin, the mid-ocean ridge would remain the weakest link between two plates, resulting in continued sea-floor spreading. If a young passive margin moves over a mantle plume, the hot spot may provide a temperature anomaly large enough to initiate re-rifting along the inner flank of the margin.

EASTERN AUSTRALIA–TASMAN SEA

The Phanerozoic history of eastern Australia records several episodes of rifting and subduction that led to terrane dispersion and amalgamation. The so-called Tasman Line (Figure 1) represents the boundary between the Proterozoic Australian Craton and different accreted terranes that might have formed in similar fashion to the North American Cordillera. The remarkable evolution of the Tasman belt

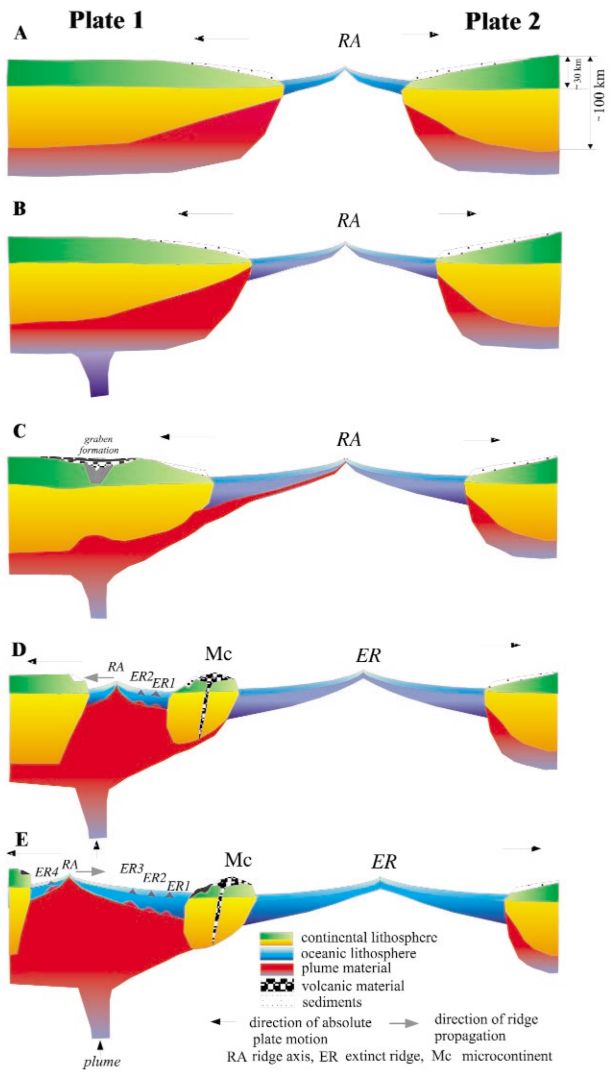

Figure 2 Conceptual model for plume-related microcontinent formation. Fragmentation of two large plates has caused a young ocean basin to form (A). A young passive margin moves into the vicinity of an existing mantle plume, enhancing the yield-strength minimum landward of the rifted margin by conductive heating (B, C). This results in renewed rifting, a jump of the active ridge towards the hot spot, and the isolation of the microcontinent (D). Further ridge jumps towards the hot spot cause subsequent asymmetries in crustal accretion, resulting in excess accretion and a series of extinct ridges on the plate including the microcontinent, if the hot spot remains under the opposite ridge flank (D). If the ridge crosses the hot spot, the sign of asymmetries in crustal accretion would switch, causing excess accretion on the ridge flank opposite the microcontinent, as in the case of Jan Mayen in the North Atlantic (E).

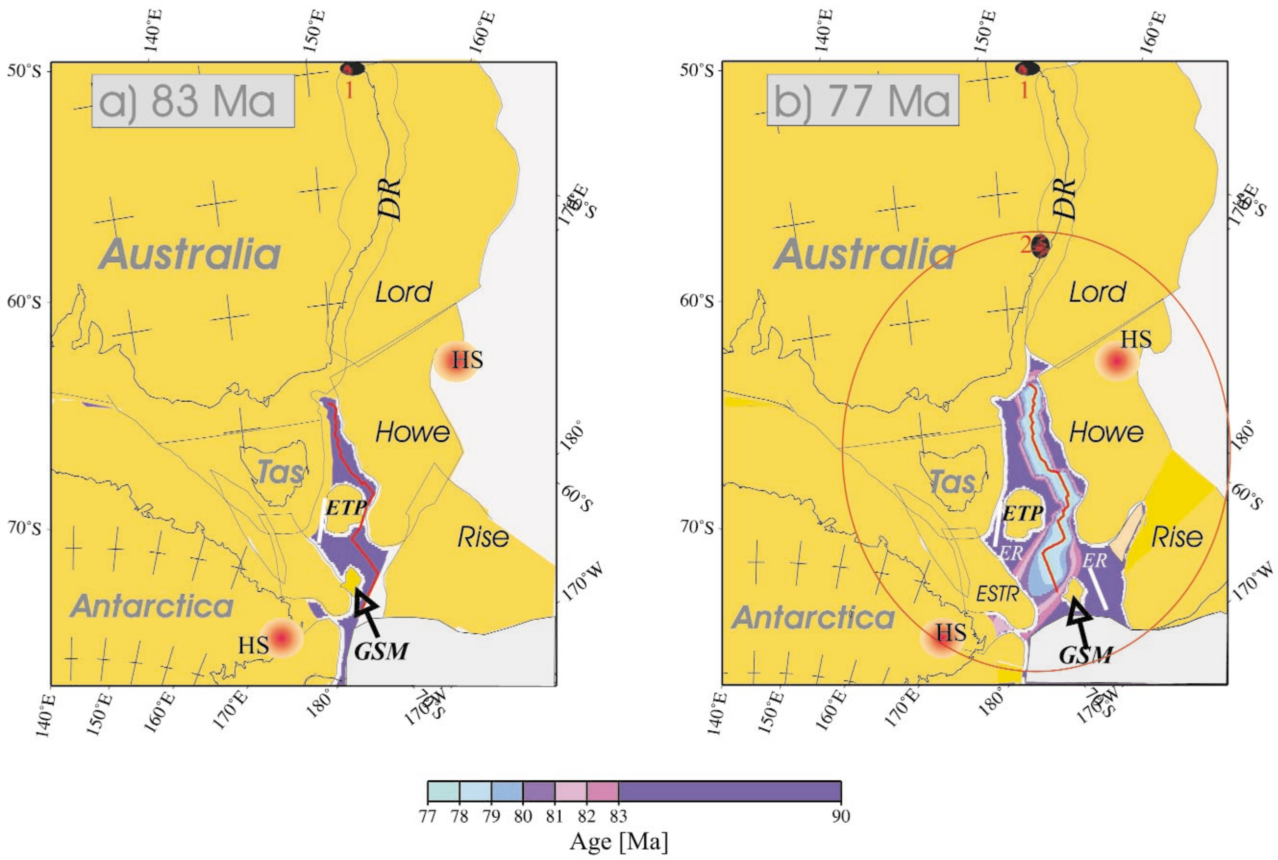

Figure 3 Reconstructed continental blocks and sea floor in the Tasman Sea at (a) 83 and (b) 77 Ma. The position of the two hot spots (HS) that were supposed to trigger ridge jumps in opposite directions at about 83 and 77 Ma are shown as large red dots. As a result two microcontinents [Gilbert Seamount (GSM) and East Tasman Plateau (ETP)] were formed in the Tasman Sea. Tas, Tasmania; ESTR, eastern South Tasman Rise; DR, Dampier Ridge. Volcanic provinces on the eastern Australian margin are shown in red/black pattern: 1, mid–Late Cretaceous Volcanic Province (ca 130–95 Ma: Bryan *et al.* 1997); 2, Sydney–Bowen Basin (118–83 Ma: Lackie & Schmidt 1996) and Barrington volcano (59 Ma: Sutherland & Fanning 2001). The red circle [in (b)] shows the approximate extent of a hot-spot swell that might have caused volcanism and microcontinent separation.

(whose components are mainly formed by oceanic crust and intra-oceanic arcs) led Coney (1992) to propose a 'quasi-continental' structure for the eastern Australian terrane. Continental breakup east of the Australian Craton started in the Late Neoproterozoic and subsequent Early Phanerozoic convergence led to terrane accretion and orogeny (Crawford *et al.* 1997; Scheibner & Basden 1998). Two examples of microcontinent formation east of Australia during the Cambrian were described by Scheibner and Veevers (2000). They suggested that at about 560–544 Ma two continental blocks (the Victorian and Gnalta microcontinents) were detached from the eastern Australian margin by the westward jump of a Palaeo-Pacific ridge and the opening of the Kanmantoo marginal basin. Scheibner and Veevers (2000) also reported that this breakup was preceded by volcanism (Mt. Arrowsmith and Glenelg volcanics). The two continental blocks were subsequently accreted to the eastern Australian continent due to the eastward subduction of the marginal basin. Crawford and Direen (1998) and Crawford and Berry (1992) also reported rifting of a continental block or microcontinental ribbon from Tasmania in the presence of a mantle plume during Late Neoproterozoic. Adams *et al.* (1998) and Pickard *et al.* (2000) suggested that continental block detachment continued in the Late Permian to Cretaceous as a suspect terrane from the southern island of New Zealand (Torlesse) travelled several hundred kilometres southward from its northeast Australian origin.

Since the Late Cretaceous the eastern Australian margin has been subjected to extension that led to the opening of a series of basins, some of them floored by extended continental crust (New Caledonia Basin, Lord Howe and Middleton Basins), others by ocean crust (Tasman Sea, Cato Trough, South Loyalty Basin). Although there are no direct evidence of the age of some of these basins (New Caledonia Basin, Lord Howe and Middleton Basins), seismic data and stratigraphic correlation indicate that they formed either at the same time, or become younger toward the Australian continent (Collot *et al.* 1987; Gaina *et al.* 1998; Kroenke 1984; Shaw 1978; Van de Beuque *et al.* 1998; Wilcox *et al.* 1980). The age of the South Loyalty Basin is still controversial: on the basis of new seismic data Auzende *et al.* (2000) suggested an Early to Late Cretaceous age that is supported by the dating of the ophiolite material (120 Ma: Prinzhofer & Berger 1981), but Crawford *et al.*(2002) question this interpretation and propose an age closer to the Tasman Sea opening (i.e. ca 85 Ma).

It has been suggested that the rift propagated from south to north in the Tasman Sea and during this process a series of ridge jumps detached continental blocks from the Australian margin or from the Lord Howe Rise and left them stranded in the Tasman Sea (Gaina *et al.* 1998; Symonds *et al.* 1988). Gaina *et al.* (2000) and Müller *et al.* (2001) suggested that two of these microcontinents (the Gilbert Seamount and East Tasman Plateau) formed due to ridge–hot-spot interactions. The East Tasman Plateau is a circular plateau east of Tasmania surrounded by oceanic crust and underlain by continental basement rocks (Exon *et al.* 1997) (Figure 1). The continental nature of the Gilbert Seamount Complex southwest of the Challenger Plateau in the southeastern Tasman Sea (Figure 1) was first suggested by Ringis (1972) and is supported by its low-amplitude magnetic anomalies and bounding normal faults (Gaina *et al.* 1998).

Both continental fragments formed by ridge jumps onto adjacent continental margins after sea-floor spreading in the southern Tasman Sea had commenced. The East Tasman Plateau was separated from the Lord Howe Rise at about chron 34 (83 Ma) and the Gilbert Seamount Complex rifted off the South Tasman Rise at roughly 77 Ma, by ridge jumps in opposing directions (Figure 3). On the East Tasman Plateau, both the event that formed the microcontinent, as well as subsequent asymmetries in crustal accretion, were due to ridge jumps toward the northeast, whereas the Gilbert Seamount Complex and its associated spreading asymmetries were formed by ridge jumps toward the southwest (Gaina *et al.* 2000).

The formation of the East Tasman Plateau (Figure 3) may be associated with the hot spot that formed the Tasmanian volcanic chain (Gaina *et al.* 2000). This hot spot would have been located under the central Lord Howe Rise at Tasman Sea breakup time (ca 95 Ma) (Gaina *et al.* 2000) as indicated by the rhyolites and tuffs recovered from the Deep Sea Drilling Project Site 207 on the central Lord Howe Rise which are dated at ca 94 Ma (McDougall & Van der Lingen 1974). A thermal anomaly in the Ross Sea area in the middle–Late Cretaceous (Stern & ten Brink 1989; Storey *et al.* 1999; Weaver *et al.* 1994) would have been located south of the South Tasman Rise (Figure 3) and could have given rise to ridge jumps toward the southeast in the southernmost Tasman Sea, creating the Gilbert Seamount Complex continental fragment, as well as subsequent excess accretion on the southern Lord Howe Rise.

The formation of microcontinents as proposed by Müller *et al.* (2001) is associated with volcanism that usually follows the separation of the continental sliver. Volcanism on the East Tasman Plateau is dated at 36 Ma (Duncan & McDougall 1989) substantially younger than its isolation as a microcontinent, and the ages of the seamounts associated with the Gilbert Seamount Complex are unknown. The lack of substantial volcanism in all four cases is probably due to rifting above a plume stem, rather than a plume head.

Other continental features east of Australia, like the Lord Howe Rise and New Caledonia–Norfolk Ridge are thousands of kilometres long and several hundreds of kilometres wide and may also considered to be microcontinents. Audley-Charles *et al.* (1988) and Metcalfe (1988) suggested that the mechanism of separation of these continental slivers may be analogous to the detachment of continental blocks and slivers in southeast Asia, therefore implying a strike-slip component. A strike-slip mechanism has been proposed to account for the narrow Australian margin that borders the Tasman Sea, the presence of the Torlesse terrane with a northeast Australian origin (Adams *et al.* 1998), and as a precursor to sea-floor spreading in the south Tasman Sea (Powell 1996). However, the repeated westward rift/ridge jumps during the formation of the New Caledonia, Middleton, Lord Howe and northern Tasman Sea Basins that led to the separation of New Caledonia–Norfolk Ridge, Lord Howe Rise and Dampier Ridge could also have been triggered by a thermal anomaly present on the eastern Australian margin between 100 and 60 Ma. Volcanism was still active around 100 Ma in the Whitsunday Volcanic Province in Queensland (130– 95 Ma) (Bryan *et al.* 1997). Other volcanic provinces formed later (the Rockhampton volcanic province dated 79–73 Ma: Sutherland *et al.* 1996), and a pronounced magnetisation event associated with uplift and erosion was recorded in the Sydney–Bowen Basin (118–83 Ma) as a result of a thermal event (Lackie & Schmidt 1996) (Figure 3). These events indicate the presence of a thermal anomaly underneath eastern Australia in the Late Cretaceous. In addition, Early Tertiary volcanic events have been recorded closed to the Barrington region in New South Wales (Sutherland & Fanning 2001) (Figure 3). K–Ar and zircon fission track dating established that the Barrington shield volcano is 59–55 Ma and might have a plume origin as its location is intersected by a north–south plume path along the Tasman margin (Sutherland & Fanning 2001). This age roughly coincides with the time of stepwise ridge propagation in the north Tasman Sea that led to the separation of the Dampier Ridge continental blocks from the Australian margin (Gaina *et al.* 1998).

Whether the formation of the Norfolk Ridge, northern Lord Howe Rise and the four fragments that make up the Dampier Ridge followed the same mechanism as proposed by Müller *et al.* (2001) depends on the nature of the crust that floors the New Caledonia, Lord Howe and Middleton Basins (Figure 1). Although there is no direct evidence of oceanic crust in the New Caledonia Basin, and most authors describe this basin as floored by extended continental crust (Eade 1988; Uruski & Wood 1991), Sutherland (1999) proposed that sea-floor spreading started at chron 34 (83 Ma) and continued during chron 33r as observed from a central linear negative magnetic anomaly. The rifting episode that preceded sea-floor spreading was associated with igneous activity that produced the positive magnetic signature described by Davy (1991). In this case, at about chron 33y (79 Ma) this ridge jumped westward to the Lord Howe and Middleton Basins, but only for short time, since there is no conclusive evidence for sea-floor spreading in these two small basins east of the Tasman Sea. Most authors describe them as extended continental crust (Gaina *et al.* 1998; Ringis 1972) although there have been few attempts to interpret oceanic magnetic anomalies in these basins (Shaw 1978). The ridge jumped again westward around 68 Ma detaching the Dampier Ridge tectonic blocks from the Australian continent (Gaina *et al.* 1998).

The wealth of continental blocks and slivers presently located east of Australia formed in the Late Cretaceous–Early Tertiary, and we interpret them as being formed largely due to ridge–plume interactions. The region appears to have been scattered by thermal anomalies that generated volcanism,

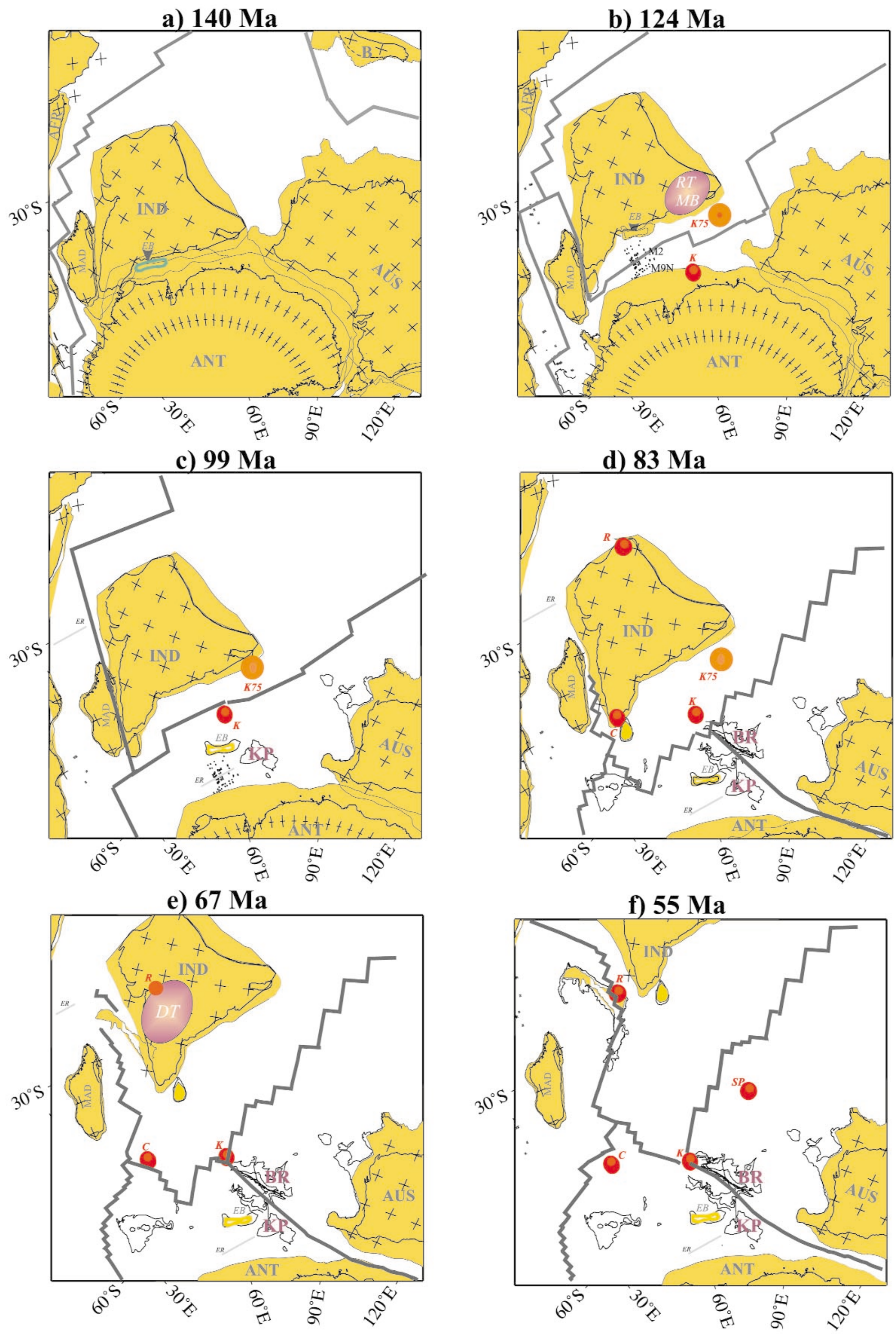

Figure 4 Indian Ocean plate reconstructions based on Kent *et al.* (in press) and revised reconstructions for the Enderby Basin. B, Burma; MAD, Madagascar; AFR, Africa; IND, India; AUS, Australia; ANT, Antarctica; RT/MB, Rajmahal Traps and Mahanadi Basin; S, Seychelles; DT, Deccan Traps; EB, Elan Bank; KP, Kerguelen Plateau; BR, Broken Ridge; K, Kerguelen hot spot; K75, Kerguelen hot spot position at 75 Ma; SP, St. Paul hot spot; C, Crozet hot spot; R, Réunion hot spot; ER, extinct ridge; M2, M9N, magnetic anomalies.

uplift and repeated rifting and sea-floor spreading. Whether they were long-lived mantle plumes, shallow mantle diapirs or whether their origin was subduction-related is unclear. Hodder (1988) proposed that the southwest Pacific region has been underlain at least during the Cenozoic by a large-scale mantle anomaly rather than by discrete plumes. On the basis of geochemical and chronological data Sun *et al.* (1991) concluded that the east Australian intraplate volcanism could be interpreted in terms of a model of the interaction of mantle plumes, shallow asthenosphere and the lithosphere. This volcanism could have been generated by a single large thermal anomaly or by several small plumes (Sutherland 1994; Wellman & McDougall 1974). Gaina *et al.* (2000) suggested that, apart from the Tasmantid seamounts, the origin of volcanism in the Tasman Sea and surrounding regions (New Zealand, Antarctica, Balleny Islands) probably resides in the upper mantle (Lanyon *et al.* 1993), rather than being generated by deep-rooted hot spots. Most of the volcanism in the Tasman Sea and south of the Tasman Sea (New Zealand, Antarctica) matches superswell-type upper mantle sources: i.e. the hot-spot tracks are short-lived and seamounts have a similar geochemistry to Pacific seamounts. This suggests that the upper mantle underlying the Tasman Sea might include a remainder of a massive upwelling event that created 'superswell'-type volcanism (Figure 3) as described by McNutt (1998) in the central Pacific.

WESTERN AUSTRALIA–INDIAN OCEAN

The western Australian margin has formed in the Late Jurassic–Early Cretaceous through rifting that propagated from northwest to southwest. During the opening of the Argo Abyssal Plain (Figure 1), a few slivers of continental crust (among them West Burma) were detached from northeast Australia and transported toward Sundaland by the opening of the Ceno-Tethys (Metcalfe 1996). The mechanism of these microcontinent-formation events might have been related to subduction of the Meso-Tethys spreading ridge under the Sundaland margin, but no geodynamic models have been proposed to show how this was accomplished.

Apart from the continental blocks that were detached from Gondwana in the Late Jurassic–Early Cretaceous, and which are now part of southeast Asia, microcontinents are relative sparse in the Indian Ocean. Examples of today's microcontinents are the Seychelles (Figure 1), and Elan Bank on the Kerguelen Plateau (Figure 1), which has also recently been shown to possess continental components (Frey *et al.* 2000; Weis *et al.* 2001). The Wallaby Plateau west of Australia has been described as a submarine volcanic plateau (Colwell *et al.* 1994), but represents a good microcontinent candidate, as seismic reflection data indicate the presence of continental material and its formation has been associated with intense volcanism and ridge jumps (see discussion in Brown *et al.* 2002). Regardless of its origin, Mihut and Müller (1998) showed that ridge–plume interaction caused two consecutive ridge propagation events transferring the Wallaby Plateau from the Indian to the Australian plate.

The Seychelles were separated from India when the Deccan–Réunion hot spot initiated sea-floor spreading along the northern Carlsberg Ridge at chron 27 (61 Ma), after spreading in the Mascarene Basin became extinct (Masson 1984; Schlich 1982) (Figure 4). Sea-floor spreading isochrons document large asymmetries in crustal accretion between the Seychelles and India (Dyment 1998; Müller *et al.* 1998). Dyment (1998) proposed that major spreading asymmetries southwest of India were caused by propagation of the ridge towards the Réunion hot spot, which was initially located east of the ridge, but to the southeast after chron 24 (52 Ma), associated with India's rapid northward motion.

Recent drilling on the Kerguelen Plateau has revealed that part of this large igneous province has continental origins (Frey *et al.* 2000; Weis *et al.* 2001). A pre-breakup reconstruction of Gondwana places Elan Bank, the continental block now part of the Kerguelen Plateau, between India and East Antarctica (Figure 4). It follows that this microcontinent was probably part of the Indian plate and transferred to the Eastern Antarctic plate by a ridge jump. In order to decipher the sequence of events that may have led to the separation of this microcontinent, we analysed recently collected geophysical data in the Enderby Basin located south of the Kerguelen Plateau (Figure 5).

Enderby Basin

The lack of marine geophysical data in the Enderby Basin southwest of the Kerguelen Plateau had led to the suggestion that the rifting process that started in northwest Australia in the Late Jurassic propagated southward and separated India from Australia and East Antarctica at about 120 Ma (Müller *et al.* 2000; Royer & Coffin 1992). As a consequence, the Enderby Basin has been considered a late Early Cretaceous basin, floored by oceanic crust formed in the Cretaceous Quiet Zone period.

New magnetic, gravity, bathymetry and seismic data collected in the Enderby Basin region (Joshima *et al.* 2001) offer the opportunity to re-examine the early sea-floor spreading history between East Antarctica and India, and to refine Mesozoic plate reconstructions in this area. We have identified a Mesozoic magnetic anomaly sequence southwest of Kerguelen Plateau off Enderby Land that ranges from anomaly M2 (126.7 Ma) to M9N (129.5 Ma) based on Gradstein *et al.*'s (1994) time-scale (Figures 5, 6). Our model assumes that the spreading rate was about 40 mm/y from the onset of sea-floor spreading (about 130 Ma) to chron M4 (126.7 Ma) and decreased to 16 mm/y until it ceased at about 124 Ma (Figure 6). Both magnetic anomaly identifications and fracture zones (Sandwell & Smith 1997) indicate that sea-floor spreading occurred in a north-northwest–south-southeast direction. The extinct spreading ridge is not very well visible on the gravity anomaly grid, but it seems to be located within a positive gravity anomaly (Figure 5). Although sea floor magnetic lineations are sparse and obscured by the Kerguelen Plateau volcanic signature north of the interpreted extinct spreading ridge, we were able to identify the Indian counterpart of the Antarctic Mesozoic anomaly sequence. As suggested by Nogi *et al.* (1992, 1996) and contrary to Ramana *et al.*'s (1994) interpretation, most of the conjugate Mesozoic anomaly sequence (from M9N to M2) can be found in the Enderby Basin. This implies that the youngest M-anomalies (M0 and M1) are either north of Elan Bank or

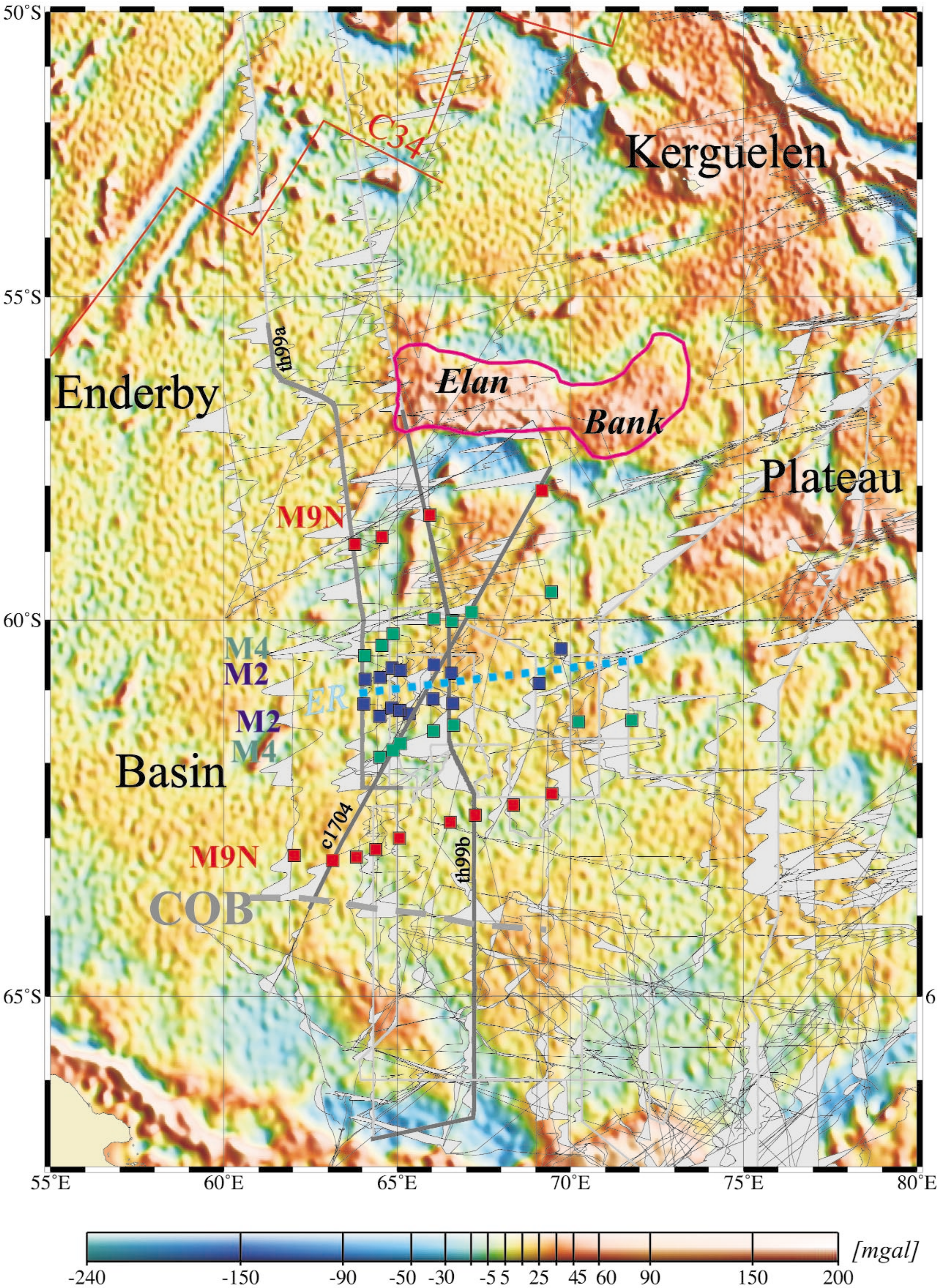

Figure 5 Satellite-derived gravity anomalies (Sandwell & Smith 1997) and magnetic anomalies along cruise tracks in the Enderby Basin. Coloured squares are magnetic anomaly identifications (M2, blue; M4, green; M9N, red). The isochron for chron 34 (83 Ma) from Müller et al's (1997) age grid is shown in red (C34). The interpreted continent–ocean boundary (COB) is the thick dashed grey line. ER, extinct ridge (light blue). The cruise tracks shown in Figure 6 are outlined in dark grey.

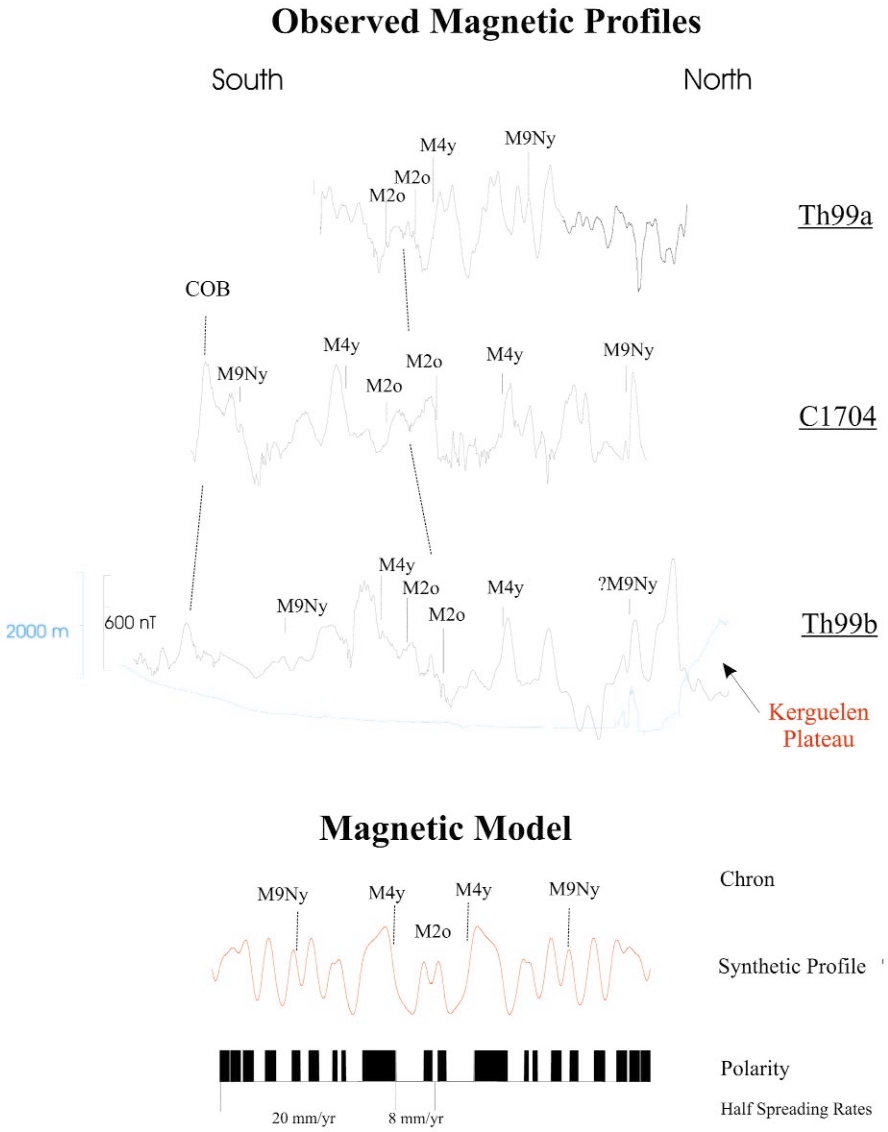

Figure 6 Selected magnetic profiles and synthetic model for the magnetic sequence identified in the Enderby Basin (see Figure 5 for location).

in the Bay of Bengal. The magnetic data north of Elan Bank are insufficient for an interpretation of the missing anomaly sequence, whereas the magnetic anomaly amplitudes in the Bay of Bengal are very low, due to the large sediment thickness. Close to the Antarctic (Enderby Land) margin, we observe a prominent high-amplitude magnetic anomaly in several profiles, followed by a low-amplitude magnetic anomaly to the south. The trend of this anomaly is oblique to all other Enderby Basin magnetic lineations, and bears some resemblance with the East Coast Magnetic Anomaly off the eastern North American margin (Klitgord & Schouten 1986). We speculate that this is the magnetic signature of the boundary between continental and oceanic crust (Figure 5), possibly corresponding to a landward edge of reversely magnetised marginal oceanic crust or intrusions generated in the transitional phase between rifting and sea-floor spreading (Klitgord & Schouten 1986).

According to this new interpretation, rifting and sea-floor spreading in the Enderby Basin was contemporaneous with the opening of the Perth Abyssal Plain southwest of Australia starting at about 130 Ma (Figure 4). Our magnetic anomaly interpretation suggests that sea-floor spreading ceased in the Enderby Basin at about 124 Ma, after the formation of magnetic anomaly M2 (124.7) (Table 1). Afterward, the mid-ocean ridge jumped onto the eastern Indian margin, separating Elan Bank as a microcontinent, and leaving two conjugate M-sequences in the Enderby Basin. The cessation of spreading in the Somali Basin at 120 Ma resulted in the amalgamation of Madagascar and Africa into one plate, and the onset of right-lateral strike-slip between Madagascar and India that continued until spreading in the Mascarene Basin between the two plates started at about 86 Ma.

Table 1 Finite rotation poles for India–East Antarctica (fixed).

Chron (Ma)	Latitude	Longitude	Angle (°)
M2 (124.7)	18.02°N	164.67°E	0.55
M4 (126.7)	38.50°N	172.21°W	2.33
M9N (129.5)	38.32°N	169.27°W	7.06
132.1 (pre-breakup)	27.32°N	178.18°W	9.82

An unresolved problem in these reconstructions is the relationship between the Kerguelen hot spot and the formation of the Elan Bank microcontinent. The northward ridge jump following magnetic anomaly M2, which presumably separated Elan Bank from the Indian margin, was likely triggered by ridge–hot-spot interaction, as is the case for many large ridge jumps and the formation of other microcontinents (Müller *et al.* 2001). However, a fixed Kerguelen hot spot would have been about 1000 km south of Elan Bank at 124 Ma (Figure 4). This discrepancy can be resolved if we take into account that hot spots were not fixed relative to each other and the Earth's spin axis (Steinberger 2000). Preliminary palaeomagnetic results from the central and northern Kerguelen Plateau (ODP Leg 183, Sites 1138 and 1140, respectively: Steinberger 2000) indicate southward motion of the Kerguelen hot spot through time. This southward motion is also supported by a 'reconstructed' Kerguelen hot spot position for 75 Ma computed by O'Neil *et al.* (2001) (Figure 4b). Apart from explaining the separation of the Elan Bank microcontinent from India and its subsequent transfer to the Antarctic plate due to ridge–hot-spot interaction, a 'reconstructed' Kerguelen hot spot closer to India for times older than 83 Ma would better account for the formation of the late Early Cretaceous Rajmahal traps in northeast India. This volcanic province, along with the Sylhet traps and volcanics in the Bengal and Mahanadi Basins (Figure 4b), covers an area almost on the scale of the Deccan volcanic province in the western India, and was formed around 118–117 Ma (Baksi 1995; Baksi *et al.* 1987; Kent *et al.* in press).

CONCLUSIONS

The ocean basins and margins around Australia are a rich natural laboratory for studying crust–mantle interactions. Plume-related microcontinent formation is a widespread, but not well-studied phenomenon, which is illustrated in abundance in the Indian Ocean and Tasman Sea. East of the eastern Australian margin there are two small continental blocks (the East Tasman Plateau and Gilbert Seamount) that were separated in the Late Cretaceous from the Australian margin, and the Lord Howe Rise, respectively, due to ridge jumps toward thermal anomalies located beneath the Lord Howe Rise and Ross Sea. We speculate that continental slivers such as the northern Lord Howe Rise, Dampier Ridge blocks and possibly the Norfolk Ridge were also isolated due to a thermal anomaly under the eastern Australian margin in Late Cretaceous–Early Tertiary (these continental blocks could be considered microcontinents only if further evidence would demonstrate the oceanic nature of surrounding basins). These thermal anomalies might have been individual hot spots or diapirs generated by a superswell or a regional mantle heterogeneity located under the Tasman Sea. Alternatively, some older terranes could have been displaced from the eastern Australian margin due to large strike-slip motion (Adams *et al.* 1998), a mechanism proposed for the formation of narrow slivers in southeast Asia (Audley-Charles *et al.* 1988; Metcalfe 1988).

In the Indian Ocean, west of Australia, Seychelles and Elan Bank are also cases of microcontinent formation due to ridge–hot-spot interactions. We suggest that Elan Bank, a recently identified microcontinent on the central Kerguelen Plateau (Frey *et al.* 2000), was isolated from India by a mid-ocean ridge jump towards the Kerguelen hot spot, which was located farther north at 120 Ma than at present. According to the oceanic crustal ages identified in the Enderby Basin based on a new set of geophysical data, the Elan Bank microcontinent became part of the East Antarctic plate around 124 Ma. The lack of concrete evidence for a continental origin in the case of Wallaby Plateau precludes its classification as a microcontinent, although its formation was associated with volcanism and ridge jumps, and seismic data indicate the presence of some continental material.

Our plume–ridge interaction model for microcontinent formation is conceptual at present. New studies on the structure and behaviour of mantle plumes show an increasing complexity of their geometry (diapirism, tilting) and behaviour (migrating as opposed to fixed) over time periods of tens of millions of years (Cox 1999; O'Connor *et al.* 2000; O'Neil *et al.* 2001), and therefore the isolation of continental slivers and scattered thermal anomalies may also be related. In order to determine exactly under which conditions ridges propagate/jump onto thinned continental lithosphere, numerical models for the interaction between thermal mantle anomalies/rising plumes and continental lithosphere with spatially varying yield-stress envelopes are required. Such models not only have the potential to elucidate how and when microcontinents form, but will also help to better model the thermal and subsidence history of sedimentary basins and hydrocarbon resources associated with microcontinents and the narrow passive margins left behind.

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