A recipe for microcontinent formation

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ABSTRACT

Accreted slivers of continental margins are common in the geologic record, but the processes that lead to their formation are poorly understood. We observe an association of plume-related microcontinent isolation and subsequent long-term asymmetries in oceanic crustal accretion based on four recent examples: the Seychelles in the Indian Ocean, Jan Mayen in the Norwegian-Greenland Sea, and the East Tasman Plateau and the Gilbert Seamount Complex in the Tasman Sea. These microcontinents formed by rerifting of a young continental margin (<25 m.y. old) in the vicinity of a mantle-plume stem, followed by asymmetric seafloor spreading. Two-dimensional numerical stochastic basin modeling suggests that a yield-strength minimum along the landward edge of a rifted margin, thermally enhanced by heating from a mantle plume, may cause a spreading ridge to jump onto this zone of weakness. This action isolates a passive-margin segment. The association of large igneous provinces and microcontinents should be useful for identifying similar events in the geological record.

Keywords: microcontinents, plumes, crustal accretion, sea-floor spreading, terranes.

INTRODUCTION

Even though many models exist for the process of terrane accretion, we know little about how microcontinents form. Geologically recently formed microcontinents were first summarized by Vink et al. (1984). They 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 favors 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 reactivation time of extension (Negredo et al., 1995). The Steckler and ten Brink model accounts for multiphase rifting, but not for the jump or propagation of a spreading ridge into a continental margin after the onset of seafloor spreading. An existing mid-ocean ridge should always yield to extension before rerifting 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 seafloor spreading, causing a breakup unconformity (Braun and Beaumont, 1989), as postbreakup divergence between two plates is accommodated by seafloor spreading. Exceptions to this rule are given by four microcontinents: the Jan Mayen microcontinent in the Norwegian-Greenland Sea, the Seychelles in the Indian Ocean, and the East Tasman Plateau and the Gilbert Seamount Complex in the Tasman Sea (Fig. 1). In these places, an existing spreading ridge propagated or jumped onto a continental margin, severing a small segment of stretched crust from a large continent, while the existing mid-ocean ridge became extinct. This process raises the question, What mechanism halts spreading along an existing mid-ocean ridge while causing the creation of a new ridge within stretched continental crust?

RIDGE-PLUME INTERACTION

Microcontinent formation can be viewed as a large ridge-jump or propagation event. To investigate the connection between oceanic ridge-jumps and microcontinent formation, we analyzed pairs of adjacent ocean-floor isochrons on the tectonic plate, including a microcontinent and its conjugate plate, and compared their local separations with symmetric plate-motion model predictions, following Müller et al. (1998) and Gaina et al. (2000). Local deviations from symmetric spreading are determined by comparing the angular distance between two adjacent seafloor isochrons with the distance that would be expected for symmetric spreading, based on a plate model determined by magnetic anomalies and fracture zones from two conjugate plates. Our results reveal that all four microcontinent formation events (Fig. 1) were accompanied by prolonged periods of asymmetries in oceanic crustal accretion. Deficits in oceanic crustal accretion mostly occur on ridge flanks overlying one or several hotspots, as ridges jump and/or propagate repeatedly toward mantle plumes (Müller et al., 1998). To test whether both microcontinent formation and nearby asymmetries in oceanic crustal accretion are triggered by plumes, we investigated the positions of known hotspots with respect to the evolving mid-ocean ridge systems around microcontinents.

JAN MAYEN MICROCONTINENT

The Jan Mayen microcontinent (Gudlaugsson et al., 1988) in the Norwegian Greenland Sea formed when spreading between Jan Mayen and Eurasia became extinct and a new spreading center formed between Greenland and Jan Mayen at about chron 7 (Nunns, 1983; Talwani and Eldholm, 1977) (Fig. 1, A and B). Seismic refraction data demonstrate that Jan Mayen is underlain by continental crust up to ~15 km thick (Kodaira et al., 1998). Roughly between chrons 13 (33 Ma) and 7 (25 Ma), simultaneous fan-shaped spreading occurred on both sides of Jan Mayen (Nunns, 1983; Müller et al., 1997) (Fig. 1A). Nunns (1983) interpreted the propagation of a new spreading center as the result of a change in spreading direction at chron 20 (43 Ma), causing compression across the transform fault linking the Reykjanes and Aegir ridge axes. However, the observed change in spreading direction is very minor and occurred earlier than microcontinent separation. Instead, we propose that a new rift along the landward edge of the Greenland margin west of Jan Mayen began to form between 40 and 33 Ma (chrons 18 and 13) (Fig. 1A), triggered by the Iceland plume. At this time, the position of the Iceland plume stem coincided with the inner edge of the rifted margin (Lawver and Muller, 1994). This event initiated northward rift propagation, gradually separating a 500-km-long segment of rifted margin from Greenland (Fig. 1A). On the basis of a dense set of magnetic data, we constructed seafloor isochrons and assessed the asymmetry of spreading through time (Fig. 1B). Pre–chron 6 spreading between Jan Mayen and Greenland was markedly asymmetric, with excess accretion on the Jan Mayen plate (Fig. 1B). After chron 6, the pattern became reversed, with excess accretion on Greenland. This reversal in the polarity of spreading asymmetry is identical to that seen on and south of Iceland (Müller et al., 1998). From the onset of spreading between Jan Mayen and Greenland until chron 6, the ridge propagated and jumped repeatedly toward the Iceland hotspot, first isolating the Jan Mayen microcontinent and subsequently transferring small segments of oceanic crust from Greenland to Jan Mayen. After chron 6, this process continued, but with the ridge jumping eastward, as the hotspot was now located east of the ridge.

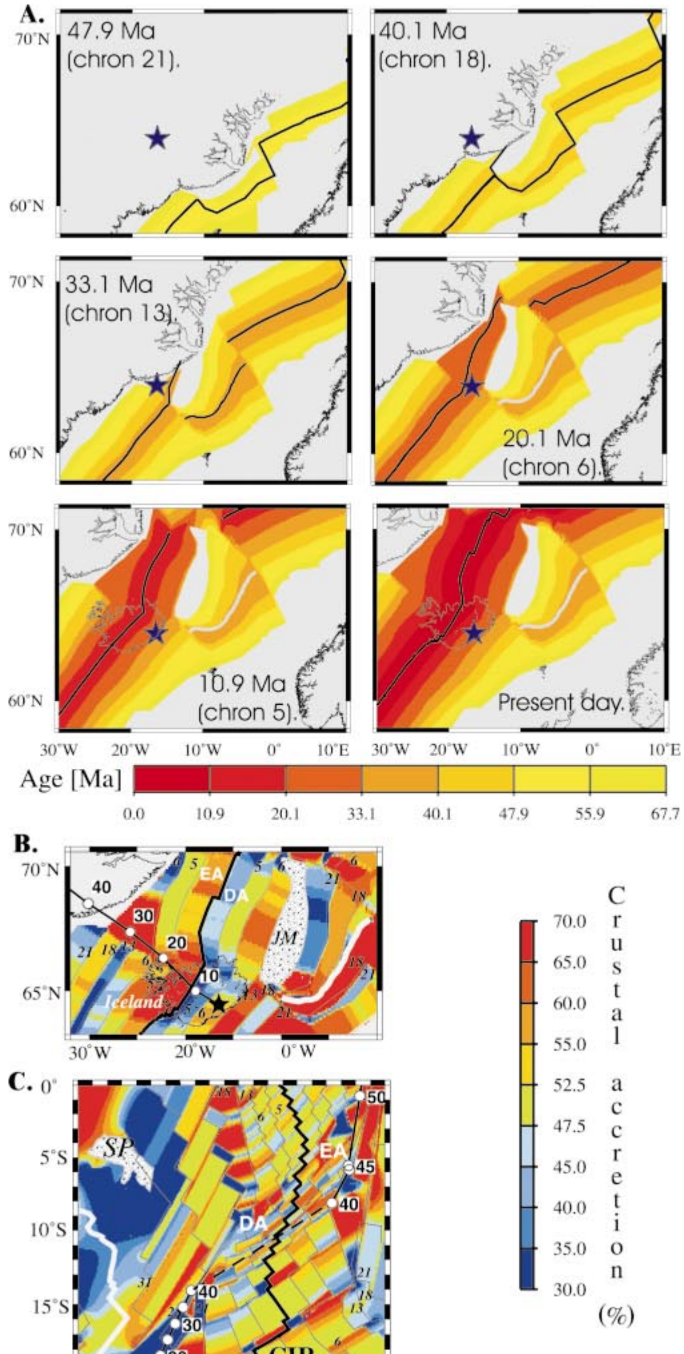
SEYCHELLES MICROCONTINENT

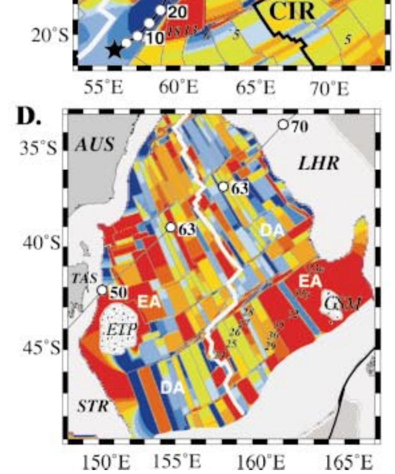
Spectacular exposures of granite make the Seychelles a type example for a microcontinent (Wegener, 1924). The Seychelles were separated from India when the Deccan-Réunion hotspot initiated seafloor spreading along the northern Carlsberg Ridge at chron 27 (61 Ma), after spreading in the Mascarene Basin stopped (Masson, 1984). Seafloor-spreading isochrons document large asymmetries in crustal accretion between the Seychelles and India (Dyment, 1998; Müller et al., 1998) (Fig. 1C). Dyment (1998) proposed that major spreading asymmetries southwest of India were caused by propagation of the ridge toward the Réunion hotspot, which was initially located east of the ridge, but to the southeast after chron 24, associated with India's rapid northward motion. Asymmetries due to ridge jumps toward the Réunion hotspot are also found along the Central Indian Ridge, with deficits in accretion of 2%–3% on the African plate since 33 Ma, when Réunion was located under the African plate, and a deficit on the Indian plate of up to 6% prior to 33 Ma when the hotspot was situated east of the ridge (Fig. 1C) (Müller et al., 1998). This sequence of events mimics that of Jan Mayen, with a young continental margin moving over a plume stem, leading to renewed rifting, microcontinent formation, seafloor spreading, and deficits in accretion on the ridge flank under which the hotspot is located, owing to ridge propagation toward the plume.

MICROCONTINENTS IN THE TASMAN SEA

The circular East Tasman Plateau is east of Tasmania and is surrounded by oceanic crust and underlain by continental basement rocks (Exon et al., 1997) (Fig. 1D). It supports the Soela Seamount guyot formed as the result of Paleogene hotspot volcanism (Duncan and McDougall, 1989). The continental nature of the Gilbert Seamount Complex southwest of the Challenger Plateau in the southeastern Tasman Sea 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 seafloor 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 about 77 Ma, by ridge jumps in opposing directions. Both events were followed by prolonged





Magnetic anomaly	Age (Ma)
5	10.9
6	20.1
13	33.1
18	40.1
21	47.9
24	53.3
25	55.9
26	57.9
27	61.2
28	62.5
29	64.0
30	65.6
31	67.7
32	71.1
33	73.6

Figure 1. A: Age grid reconstructions from Müller et al. (1997) in a hotspot reference frame from Müller et al. (1993) showing position of Iceland hotspot (star) relative to Greenland-Eurasia plate boundary. B–D: Oceanic crustal accretion rates and hotspot tracks (Iceland, Réunion, and unnamed hotspot) in Norwegian-Greenland Sea, northern central Indian Ocean, and southern Tasman Sea, respectively. Hotspot tracks are labeled in million years; star marks present-day location of hotspot. Two locations of hotspot are shown for 45 Ma in northern central Indian Ocean (C); their separation reflects recent diffuse intraplate deformation in central Indian Ocean. Asymmetries in oceanic crustal accretion are shown as percentage of crust on one ridge flank relative to total amount of crust formed in a given time interval on both sides of ridge; warm colors indicate excess accretion (EA), and cold colors indicate deficient accretion (DA). Heavy black lines are active plate boundaries; light gray lines are extinct ridges. Isochrons are labeled with chron numbers, indicating beginning or termination of periods of normal magnetic polarity. Light-gray pattern shows continental margins. SP—Seychelles Plateau, CIR—Central Indian Ridge, TAS—Tasmania, STR—South Tasman Rise, ETP—East Tasman Plateau, GSM—Gilbert Seamount Complex.

seafloor spreading asymmetries (Fig. 1D). On the East Tasman Plateau, both the event that formed the microcontinent and subsequent aymmetries 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) proposed that both microcontinents and associated spreading asymmetries were caused by ridge-hotspot interaction. The formation of the East Tasman Plateau (Fig. 1D) may be associated with the hotspot that formed the Tasmanian volcanic chain. This hotspot would have been located under the central Lord Howe Rise at Tasman Sea breakup time (ca. 95 Ma) (Gaina et al., 2000). Rhyolites and tuffs recovered from Deep Sea Drilling Project Site 207 on the central Lord Howe Rise are dated at ca. 94 Ma (McDougall and Van der Lingen, 1974). Also, Willcox et al. (1980) identified Late Cretaceous–early Tertiary intrusions in seismic data acquired from the central Lord Howe Rise, supporting the presence of a thermal anomaly under the central Lord Howe Rise during the early opening of the Tasman Sea.

Weaver et al. (1994) and Storey et al. (1999) suggested, on the basis of available data, including Pb isotopes, that the Cretaceous rift-related magmatism recorded in New Zealand and Marie Byrd Land was generated in the vicinity of a mantle plume, which controlled the position of breakup. A thermal anomaly in the Ross Sea area in the Late Cretaceous would have been located south of the South Tasman Rise and could have given rise to ridge jumps toward the southeast in the southernmost Tasman Sea, creating both the Gilbert Seamount Complex continental fragment and subsequent excess accretion on the southern Lord Howe Rise.

VOLCANISM AND MICROCONTINENT FORMATION

In all four examples investigated here, volcanism was associated with microcontinental isolation, but it was not particularly voluminous and generally postdates the rifting event. On Jan Mayen, a volcanic sequence identified from seismic reflection data postdates block faulting (Gudlaugsson et al., 1988) and is relatively thin (Kodaira et al., 1998). On the Seychelles, the emplacement of Late Cretaceous tholeiitic dikes, similar in composition to the Deccan basalts from India, was followed by minor alkalic volcanism (Devey and Stephens, 1992). The age gap between tholeiitic and alkalic volcanism, a succession typical for oceanic hotspots, is estimated at 3–5 m.y., centered on 64 Ma (Devey and Stephens, 1992). Therefore, Devey and Stephens (1992) suggested that the younger alkalic magmatism that postdates the tholeiitic magmatism observed in the Seychelles, is the result of the Réunion plume. The volcanic history of the Tasman Sea microcontinents is less well known. Volcanism on the East Tasman Plateau is dated at 36 Ma (Duncan and McDougall, 1989), substantially younger than its isolation as a microcontinent, and ages of the seamounts associated with the Gilbert Seamount Complex are unknown. The lack of substantial volcanism in all four cases is likely due to rifting above a plume stem, rather than a plume head. However, for the two primary examples, Jan Mayen and the Seychelles, the hotspot tracks associated with their formation can clearly be traced back to a large igneous province (LIP)—the North Atlantic volcanic province and the Deccan Traps, respectively (Coffin and Eldholm, 1994). This association is important for identifying plume-related microcontinents in the geologic record.

MODEL FOR PLUME-RELATED MICROCONTINENT FORMATION

We propose that a sequence of five events leads to the formation of plume-related microcontinents: (1) emplacement of a LIP above a mantle plume head, rifting, and seafloor spreading between two large plates, (2) the propagation or jump of the active ridge from the spreading axis to the landward edge of a young continental margin, leading to renewed rifting, (3) minor volcanism, (4) seafloor spreading and separation of a microcontinent, and (5) a prolonged period of asymmetric seafloor spreading (Fig. 2).

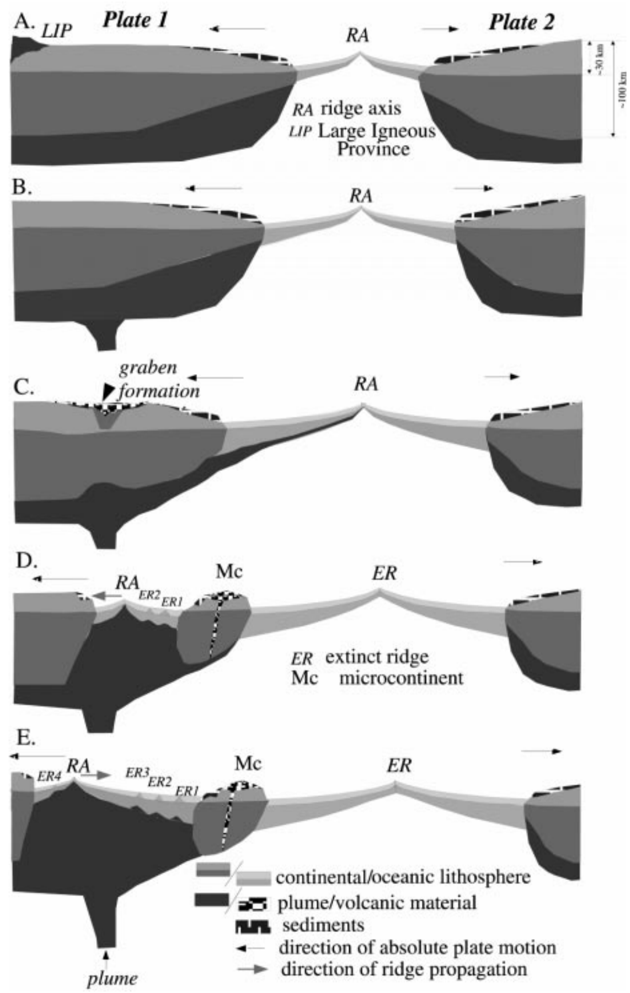


Figure 2. Conceptual model for plume-related microcontinent formation. A: Fragmentation of two large plates has caused a young ocean basin to form. B: Young passive margin moves into vicinity of existing mantle plume, enhancing yield-strength minimum landward of rifted margin by conductive heating. C,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

Steckler and ten Brink's (1986) model suggests that a yield-strength minimum occurs along the edges of a rifted margin, because there the crust is not only thickest but also hotter than normal, owing to conductive heat flow from the adjacent rift. Nielsen and Hansen (2000) suggested that heat flow deviates from the central parts of a basin toward the basin flanks, owing to thermal refraction, associated with thermal blanketing of sediments. This results in relatively low viscosities of the mantle material even a long time after basin formation (Fig. 3), facilitating reactivation of the mantle beneath the basin flanks. However, rerifting along a margin's inner edge can occur only 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 seafloor spreading. If a young passive margin moves over a

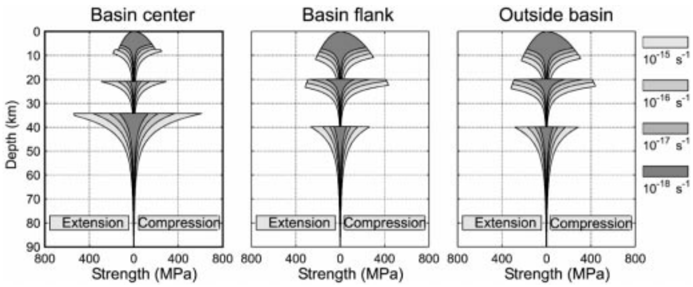


Figure 3. Lithospheric rheology from two-dimensional basin simulation (Nielsen and Hansen, 2000). Each profile is shown for four different depth-invariant strain rates. Basin center is characterized by relatively strong mantle material and relatively weak crustal material. Weakest mantle is found beneath landward edge of passive margin.

mantle plume, the hotspot may provide a temperature anomaly large enough to initiate rerifting along the inner flank of the margin. The age of the margins investigated here at the time of rerifting was 5 m.y. (East Tasman Rise), 10 m.y. (Gilbert Seamount Complex), 20 m.y. (Jan Mayen), and 25 m.y. (Seychelles). There are no examples for microcontinent formation at older margins, perhaps because they are too far away from active ridges for a ridge jump to be feasible and because they slowly regain rheologic strength.

IMPLICATIONS FOR ACCRETED TERRANES

Coney et al. (1980) identified seven fragments or slices of continental margins within the North American Cordillera, including shallow continental-margin sequences of carbonates or siliciclastic rocks in association with metavolcanic units. A westward jump of the Mid-Iapetus Ridge onto North America in the early Cambrian formed a microcontinent along the Ouachita Rift, accompanied by igneous activity (Thomas and Astini, 1996), analogous to our model. St-Onge et al. (2000) identified an apparent landward jump of an active ridge onto the northern margin of the Superior province following the northwest migration of the ''Ungava hotspot'' at ca. 2.22 Ga. Therefore, hotspot-triggered microcontinent formation may have been an important mechanism since the Proterozoic for creating continental terranes, which were later accreted. However, our model does not apply to those microplates that form in association with the transition from subduction to rifting, such as Baja California.

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