Tectonophysics 744 (2018) 403–421

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Insights from North America's failed Midcontinent Rift into the evolution of continental rifts and passive continental margins☆

Seth Stein[a,*], Carol A. Stein[b], Reece Elling[a], Jonas Kley[c], G. Randy Keller[d], Michael Wysession[e], Tyrone Rooney[f], Andrew Frederiksen[g], Robert Moucha[h]

[a] Northwestern University, United States of America
[b] University of Illinois at Chicago, United States of America
[c] Georg-August-Universität Göttingen, Germany
[d] University of Oklahoma, United States of America
[e] Washington University, United States of America
[f] Michigan State University, United States of America
[g] University of Manitoba, Canada
[h] Syracuse University, United States of America

ARTICLE INFO

Keywords:
Continental rifting
Volcanic passive margins
Midcontinent Rift

ABSTRACT

Continental rifts evolve along two possible paths. In one, a rift successfully evolves into seafloor spreading, leaving the rift structures buried beneath thick sedimentary and volcanic rocks at a passive continental margin. Alternatively, the rift fails and remains as a fossil feature within a continent. We consider insights into these processes from studies of North America's Midcontinent Rift (MCR). The MCR combines the linear geometry of a rift formed at a plate boundary and the huge igneous rock volume of a Large Igneous Province. The rift is a fault bounded basin filled with volcanics and sediments, which record a history of extension, volcanism, sedimentation, subsidence, and inversion. The MCR came close to evolving into an oceanic spreading center, but it instead failed and thus records a late stage of rifting. It thus preserves a snapshot of a stage of the process by which actively extending rifts, characterized by upwelling mantle and negative gravity anomalies, evolve either into failed and often inverted rifts without upwelling mantle and positive gravity anomalies or into passive continental margins. Many rifts can be viewed as following a generally similar evolutionary sequence, within which a complex combination of factors control the variability of structures within and among rifts. Study of the MCR also gives insight into passive continental margins. The MCR gives a snapshot of deposition of a thick, dense, and highly magnetized volcanic section during rifting. Surface exposures, seismic, and gravity data delineate a rift basin filled by inward dipping flood basalt layers, underlain by thinned and underplated crust. The fact that the MCR shows many features of a rifted volcanic margin suggests that it came close to continental breakup before it failed, and illustrates how many passive margin features form prior to breakup.

1. Introduction

Plate tectonics shapes the evolution of the continents and oceans via the Wilson cycle, in which continents rift apart to form new oceans that may grow to the size of the Atlantic and Pacific before closing and vanishing (Fig. 1). In such cases, the fate of continental rifts is to end up as passive continental margins. However, many continental rifts, such as North America's Reelfoot Rift and Southern Oklahoma Aulacogen, fail to develop into seafloor spreading centers. Such failed rifts become an important part of the fabric of the continents. As a result, failed rifts preserve "fossil" features of the rifting process that can be difficult to observe elsewhere.

In this paper, we suggest that useful insights into rift evolution can be obtained from study of one of the world's most impressive failed rifts, North America's Midcontinent Rift (MCR). The MCR is a 3000-km-long U-shaped band of buried igneous and sedimentary rocks that outcrops near Lake Superior (Figs. 2, 3). To the south, it is buried by younger sediments, but easily traced because the igneous rocks are dense and highly magnetized (King and Zietz, 1971; Hinze et al., 1997; Merino et al., 2013). The western arm extends across Minnesota and Iowa at

☆ Review Article

* Corresponding author.
E-mail address: seth@earth.northwestern.edu (S. Stein).

https://doi.org/10.1016/j.tecto.2018.07.021
Received 18 March 2018; Received in revised form 22 July 2018; Accepted 24 July 2018
Available online 29 July 2018

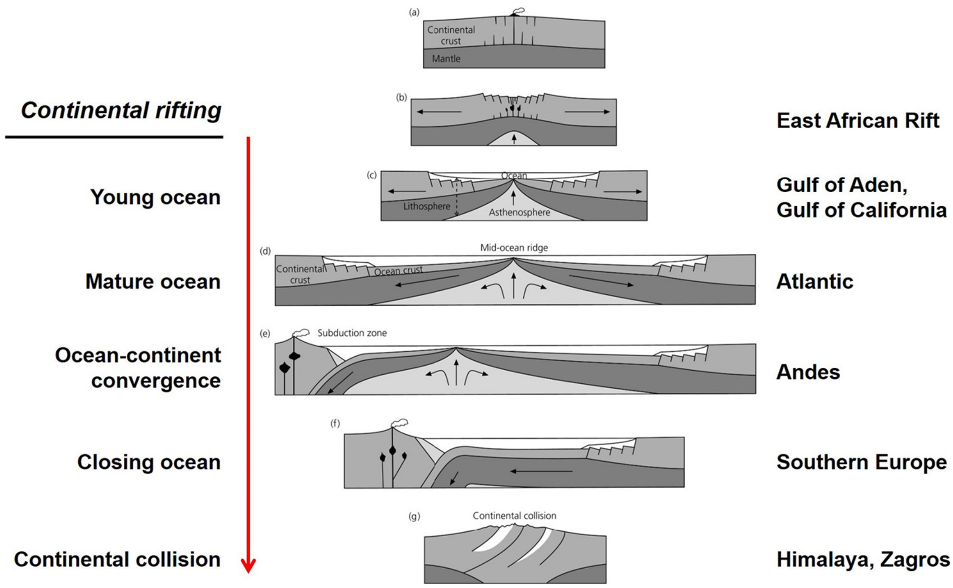

Fig. 1. Schematic illustration of the Wilson cycle, showing modern areas at each stage (Stein and Wysession, 2003).

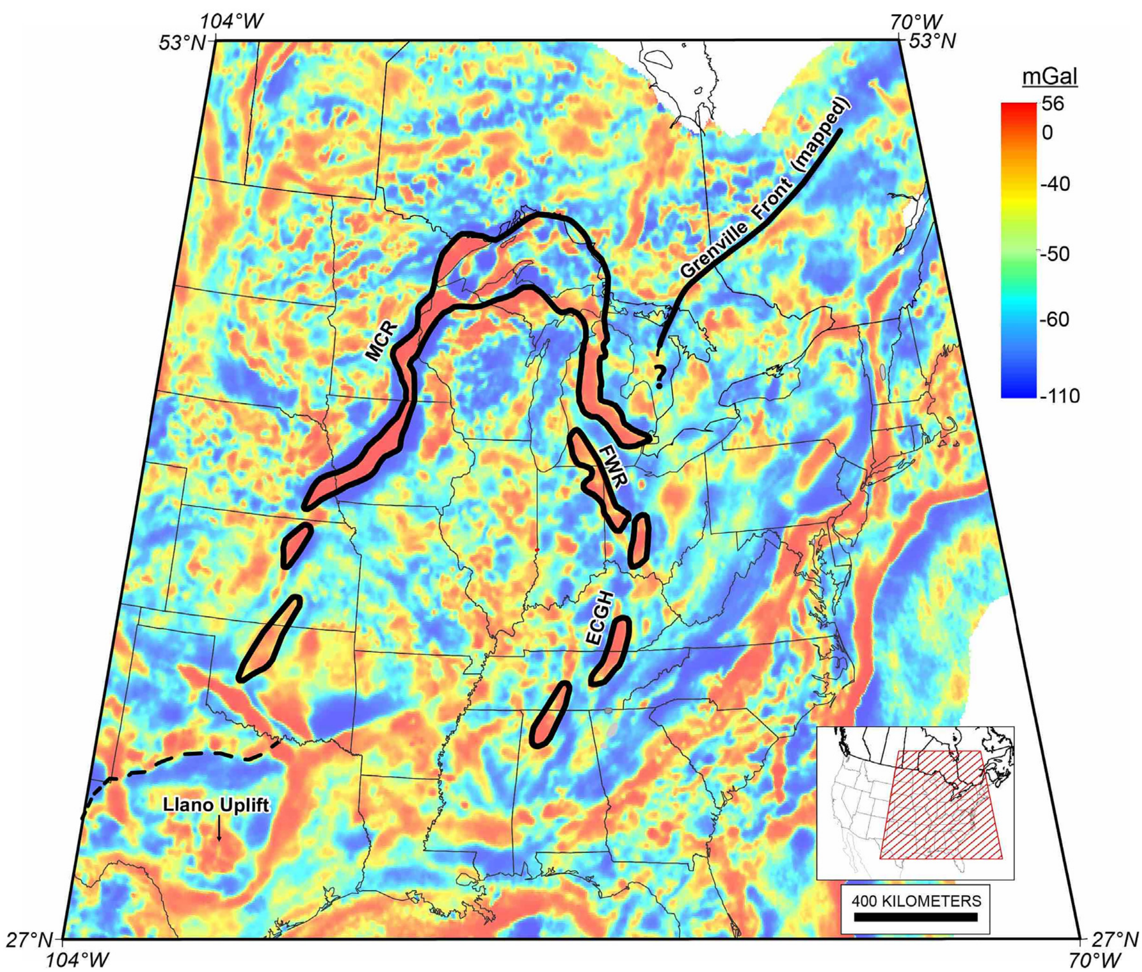

Fig. 2. Gravity map showing the MCR, including the Fort Wayne Rift (FWR) and East Continent Gravity High (ECGH) segments of the east arm, computed by upward continuing complete Bouguer anomaly (CBA) data to 40 km and subtracting result from CBA. Dashed line shows Grenville Front in Texas (Stein et al., 2015).

least to Oklahoma, and perhaps Texas and New Mexico, as evidenced by similar-age diffuse volcanism (Adams and Keller, 1994, 1996; Bright et al., 2014). The eastern arm extends southward through Michigan, Indiana, Ohio, Kentucky and Tennessee to Alabama (Keller et al., 1982; Dickas et al., 1992; Stein et al., 2014, 2018). MCR-related igneous activity may also occur near but off the arms (Drenth et al., 2015).

The MCR records a major rifting event at ~1.1 Ga during the Grenville orogeny, the sequence of events from ~1.3–0.98 Ga

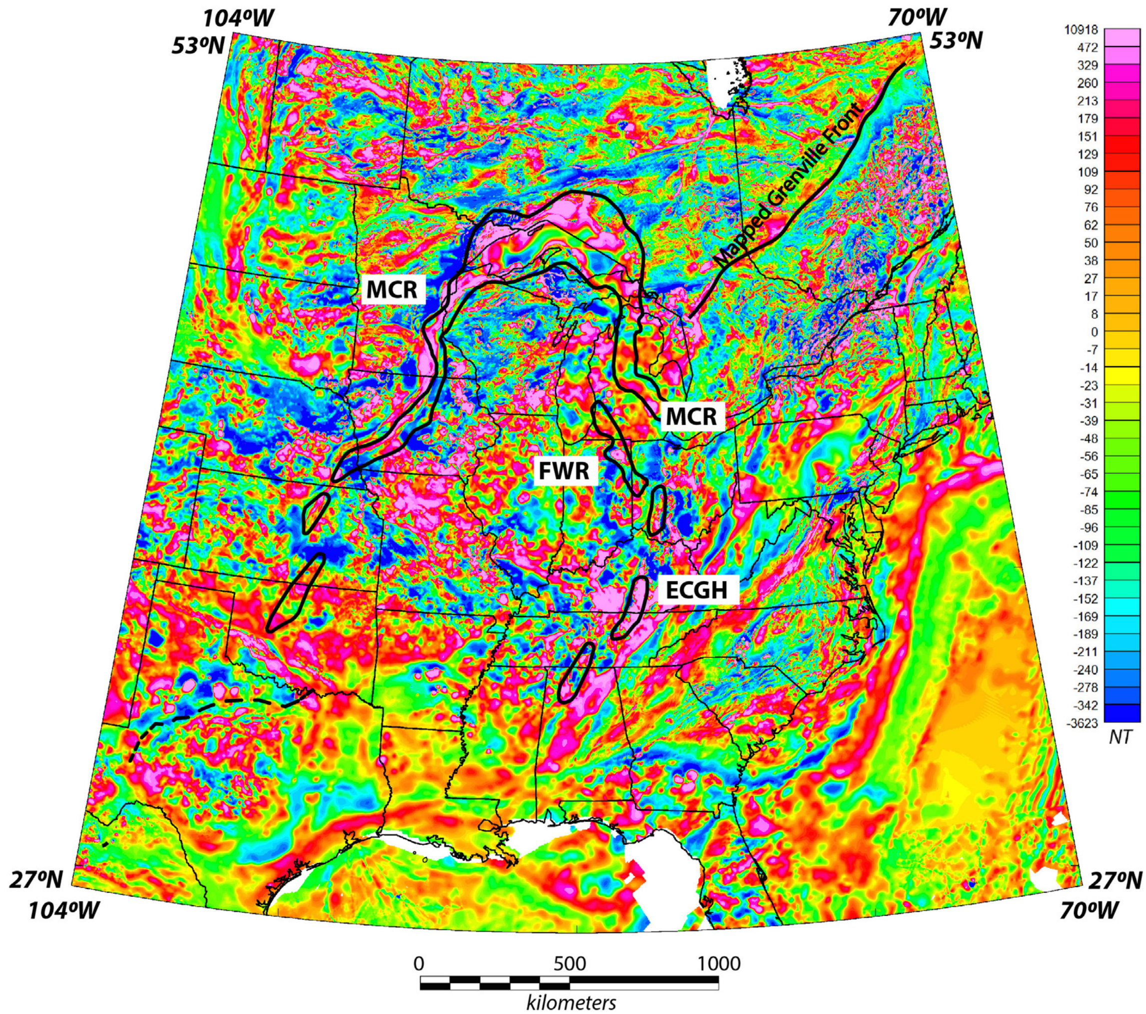

Fig. 3. Magnetic anomaly map of the region. Outlines of the Midcontinent Rift (MCR), including the Fort Wayne Rift (FWR) and East Continent Gravity High (ECGH) segments of the east arm, are from gravity data (Fig. 2). Dashed line shows Grenville Front in Texas.
Data source *https://pubs.usgs.gov/of/2002/ofr-02-414/* (Stein et al., 2018)

culminating in the assembly of a number of continental blocks into the supercontinent of Rodinia (Fig. 4) (Dalziel, 1991; Hoffman, 1991; Rivers et al., 2012; Tohver et al., 2006; Swanson-Hysell et al., 2014; Merdith et al., 2017). This rifting failed to split Laurentia, the Precambrian core of the North American continent, leaving a failed rift that did not evolve into full seafloor spreading. This rift was later inverted by regional compression, uplifting the volcanic rocks within the rift, so that some are exposed at the surface today.

How the MCR formed during the Grenville orogeny, a collisional and hence compressive series of events, is a long-standing question. Much of the question involves the MCR's relation to the Grenville Front, the continentward (western) extent of deformation of the fold and thrust belt from the Grenville orogeny (Rivers et al., 2012). As shown in Fig. 2, the Front is observed in SE Canada from surface geology and reflection seismic data, and has traditionally been assumed to extend southward into the central U.S. along the East Coast Gravity High (ECGH), a lineation of gravity and magnetic anomalies (e.g., Zietz et al., 1966; Hoffman, 1988; Whitmeyer and Karlstrom, 2007; Baranoski et al., 2009; Bartholomew and Hatcher, 2010). However, recent analysis (Stein et al., 2018) shows that these anomalies previously assumed to define the Front in the central U.S. appear to be the southward continuation of the MCR's east arm (Fig. 5). This view arises because the gravity highs along the ECGH are similar to those elsewhere along the MCR, in showing a distinct central high, presumably largely due to the dense igneous rocks filling the rift. In contrast, no similar high occurs across the Grenville Front in Canada. Moreover, many of the wells in areas of the gravity high in Ohio and Kentucky bottom in mafic rocks (Drahovzal et al., 1992; Buening, 2013) similar to MCR rocks exposed near Lake Superior and in the buried west arm (Walker and Misra, 1992; Lidiak, 1996). Hence in our view the rift basins in Ohio, Kentucky, and Indiana (Moecher et al., 2018) are part of the MCR's east arm.

Although the fact that the MCR formed during the Grenville orogeny seems surprising, the orogeny occurred in discrete compressional phases (Fig. 6). As discussed later, the MCR probably formed as part of the rifting of Amazonia from Laurentia. This occurred between compressional phases of the Grenville orogeny, prior to the Ottawan phase, a major compressional phase during which the Grenville Front (GF) formed in SE Canada.

These results offer new insights into the formation of the MCR and the Grenville Front and their association with the assembly of Rodinia (Table 1). The traditionally assumed Front's location near southeast Michigan implies that the MCR's east arm ended there, presumably because propagation of the rift extension and volcanism were stopped by the preexisting Front (Cannon et al., 1989). However, it now appears that the MCR formed before the Grenville Front. Hence the absence of an observed Front in the central U.S. may reflect its being obscured by the MCR's east arm or younger tectonic events. Alternatively, a distinct Front never formed there.

In this paper, we summarize key results about the MCR, many of

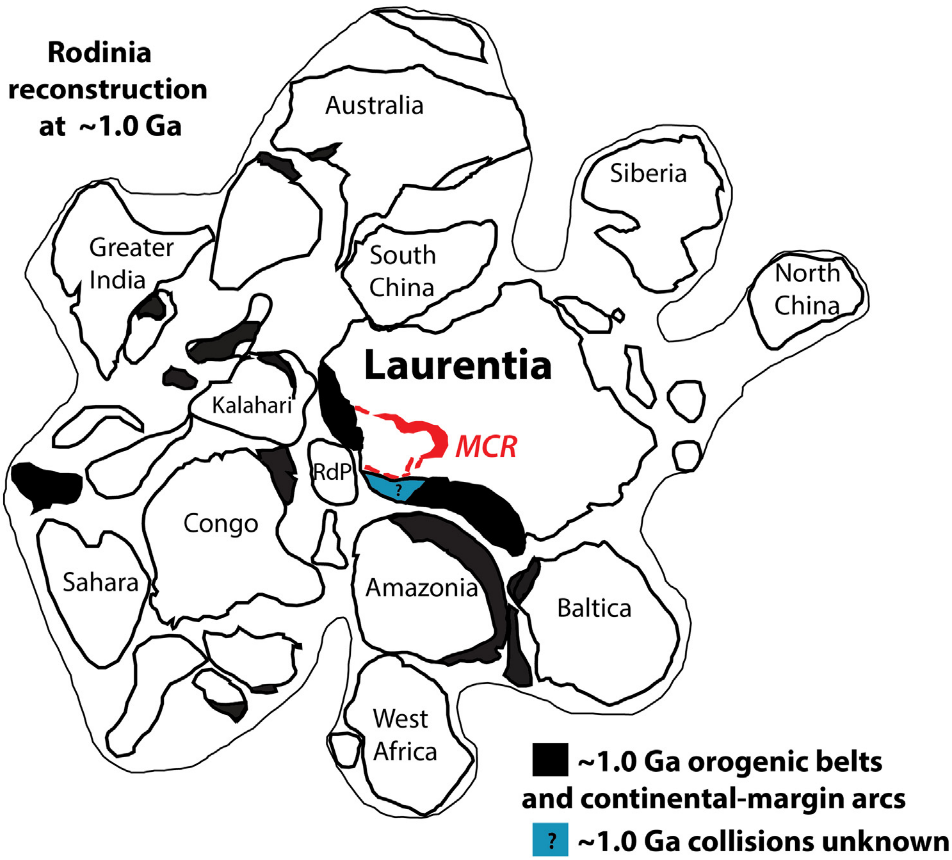

Fig. 4. Reconstruction showing major blocks and Grenville-age orogenic belts associated with the accretion of the Amazonia and Rio de la Plata (RdP) blocks to Laurentia, the core of Precambrian North America. The Grenville Front is the continentward extent of deformation of the orogenic fold and thrust belt in Laurentia (Stein et al., 2018, after Li et al., 2008). Most features shown are common to different reconstructions, but some differ owing to limitations in the paleomagnetic data available.

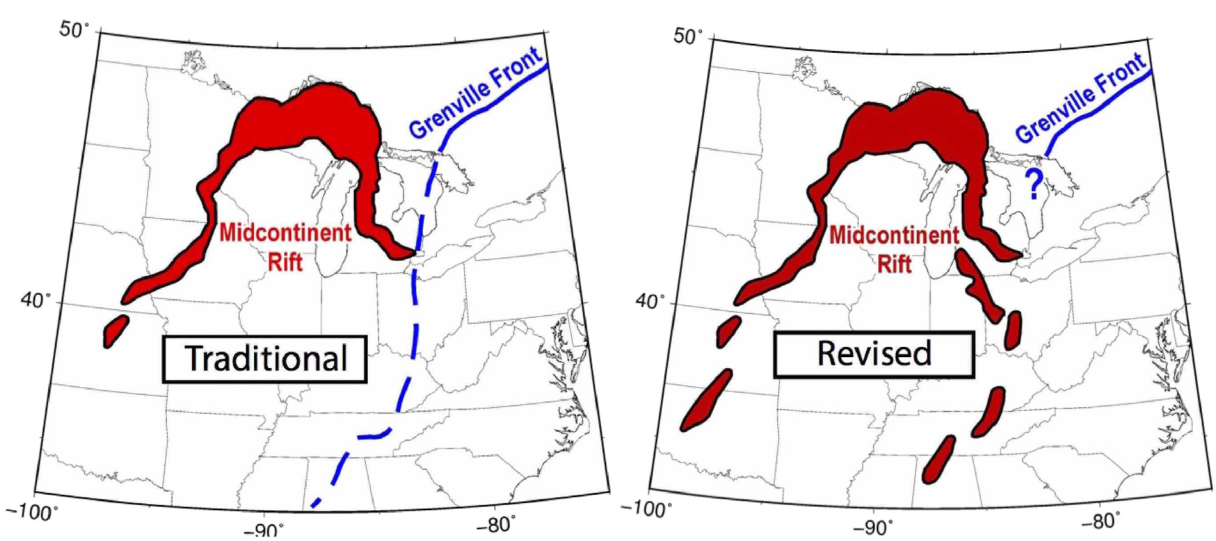

Fig. 5. (Left) Traditionally assumed geometry in which the Grenville Front truncates the east arm of the MCR and extends southward along subsurface features indicated by gravity and magnetic anomalies. (Right) Revised geometry proposed by Stein et al. (2018), in which the previously assumed Front in the central U.S. is the southward continuation of the MCR.

them from recent studies. We then use these results to provide insights into the evolution of rifts and passive margins.

2. Midcontinent Rift attributes relevant for rift evolution

The MCR has been the subject of extensive recent studies as part of the U.S. National Science Foundation's EarthScope program "*to integrate geological and geophysical data to understand the growth and modification of North America over billion-year time scales*." These studies built on and extended earlier work. Some by ourselves and coworkers involved seismological studies using USArray transportable stations and the SPREE flexible array deployment along part of the west arm (Stein et al., 2011; Wolin et al., 2015; Ola et al., 2015; Al-Eqabi et al., 2015; Zhang et al., 2016; Frederiksen et al., 2017) and integrated analysis of gravity, seismic reflection, paleomagnetic, and geological data (Merino et al., 2013; Stein et al., 2014, 2015; Malone et al., 2016) to develop models of the MCR's structure and evolution (Stein et al., 2016, 2018). Other researchers have also obtained new results (e.g., Craddock et al., 2013; Levandowski et al., 2015; Shen et al., 2013; Shen and Ritzwoller, 2016; Swanson-Hysell et al., 2014; Wunderman et al., 2018).

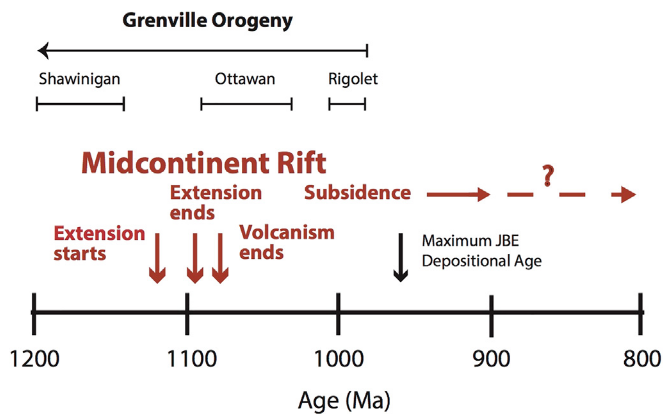

Fig. 6. Timeline for evolution of the Midcontinent Rift (MCR) and major compressional phases of the Grenville orogeny. JBE are Jacobsville, Bayfield, and other equivalent sandstones deposited above MCR post-rift sediments, which provide age control on MCR evolution. The maximum possible JBE age is derived from detrital zircon ages (Malone et al., 2016).

Table 1
Alternative models of the relation between the MCR and Grenville Front in the central U.S.

	Traditional model	*Revised model.*
Tectonic setting	MCR formed during convergence	MCR formed during Amazonia - Laurentia divergence
Temporal relationship	GF formed before MCR	MCR formed before GF
Spatial relationship	MCR rifting truncated against GF	If GF extended far enough south, GF propagation truncated against MCR or to the east

This paper summarizes key results about the MCR, most from recent studies, that provide insights into rift evolution. Although using data from a billion-year-old rift may seem strange for this purpose, it makes sense because the MCR was massively inverted and uplifted by regional compression long after it failed, so its structure is better known than failed rifts that incurred lesser degrees of inversion. It can be viewed as an end member of one of the two possible paths for the evolution of actively extending rifts. In one, a rift successfully evolves into seafloor spreading, leaving the rift structures buried beneath thick sediments at a passive continental margin. Alternatively, the rift fails, is often inverted, and is left as a fossil feature within a continent. Conceptually, our approach can be viewed as starting from the failed MCR and going backwards in time, to consider what insights we can gain about rifts at different stages in their evolution (Fig. 7).

This paper is not a comprehensive review of the MCR or of rifts and passive margins worldwide. It builds on studies of the MCR, reviewed

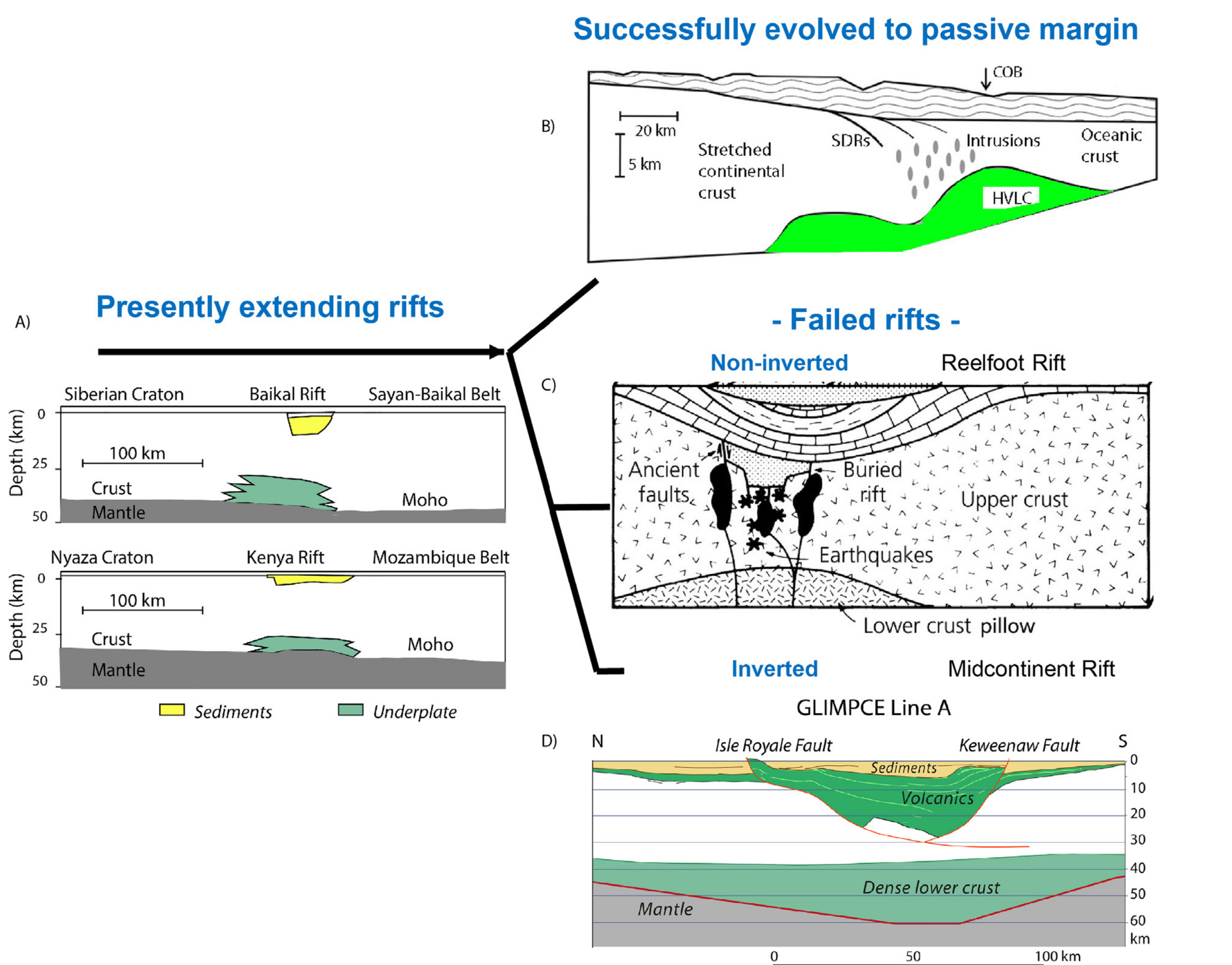

Fig. 7. Schematic sequence of rift evolution illustrated by various rifts. Panels are from A) after Thybo and Artemieva (2013), B) Schnabel et al. (2008), C) Braile et al. (1986), and D) Stein et al. (2015) modified from Green et al. (1989). COB denotes continent-ocean boundary, SDR denotes seaward dipping reflectors, and HVLC denotes high-velocity lower crustal bodies.

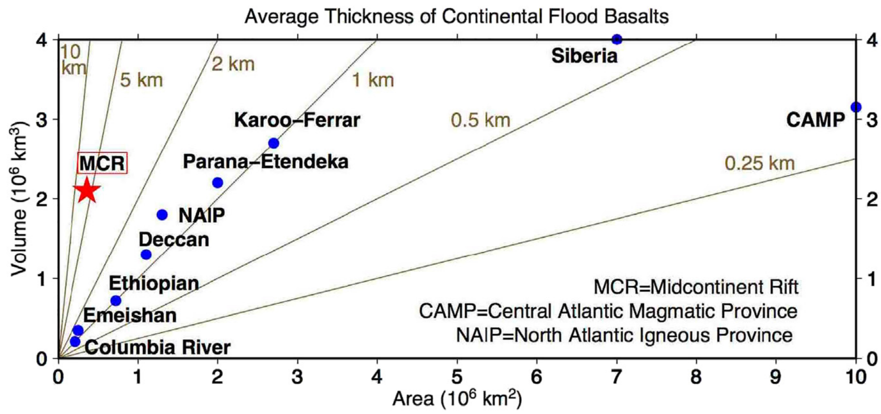

Fig. 8. Comparison of volume, surface area, and average thickness (diagonal lines) for continental flood basalts. The MCR volcanics are comparable in volume to other flood basalts, but thicker because they were deposited in a narrow subsiding rift.
After Stein et al. (2015) with new values from Ethiopia (Rooney, 2017).

by Van Schmus and Hinze (1985), Ojakangas et al. (2001), and Stein et al. (2016) and of continental rifting, reviewed by Sengör and Burke (1978), Ziegler and Cloetingh (2004), Merle (2011), Allen and Armitage (2012), Roberts and Bally (2012), Thybo and Artemieva (2013), and Frizon de Lamotte et al. (2015). As these reviews discuss, individual rifts vary significantly. For example, the presently extending Baikal rift has much less igneous fill than the MCR, and the Kenya rift (western branch of the East African Rift system) has much less crustal thinning than the MCR. Even so, consideration of the MCR can provide insights into general aspects of the rifting processes. To do this, we first summarize several key relevant features of the MCR.

2.1. A rift with extensive volcanism

Recent analysis shows that the MCR has an unusual combination of the geometry of a rift and the huge igneous rock volume of a Large Igneous Province (LIP) (Green, 1983), two types of features that differ in geometry and origin (Foulger, 2011). Rifts are segmented linear depressions, filled with sedimentary and igneous rocks, that form by extension and often evolve into plate boundaries (Roberts and Bally, 2012). Flood basalts, a class of Large Igneous Provinces, are broad regions of extensive volcanism that form due to sublithospheric processes such as a mantle plume or other mantle anomaly (Morgan, 1971, 1983; Ernst, 2014). Typical rifts are not filled with thick flood basalts, and typical continental flood basalts are not erupted in association with significant crustal extension and faulting. Modeling of MCR seismic and

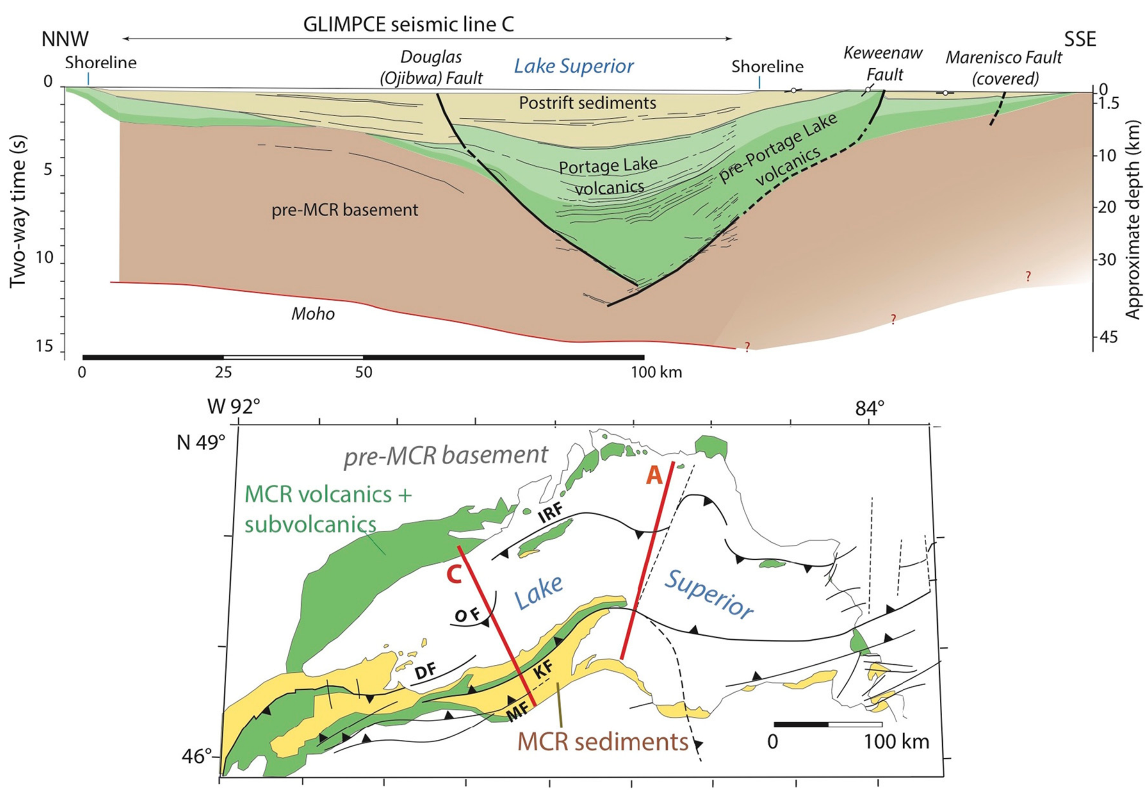

Fig. 9. MCR cross-section based on line drawing of GLIMPCE seismic line C, modified from Green et al. (1989) and complemented with land data, showing geometry of volcanic rocks and postrift sediments in the rift basin. Lower panel shows locations of lines A and C. DF is Douglas Fault, OF is Ojibwa Fault, IRF is Isle Royale, KF is Keweenaw Fault, and MF is Marenisco Fault (Stein et al., 2015).

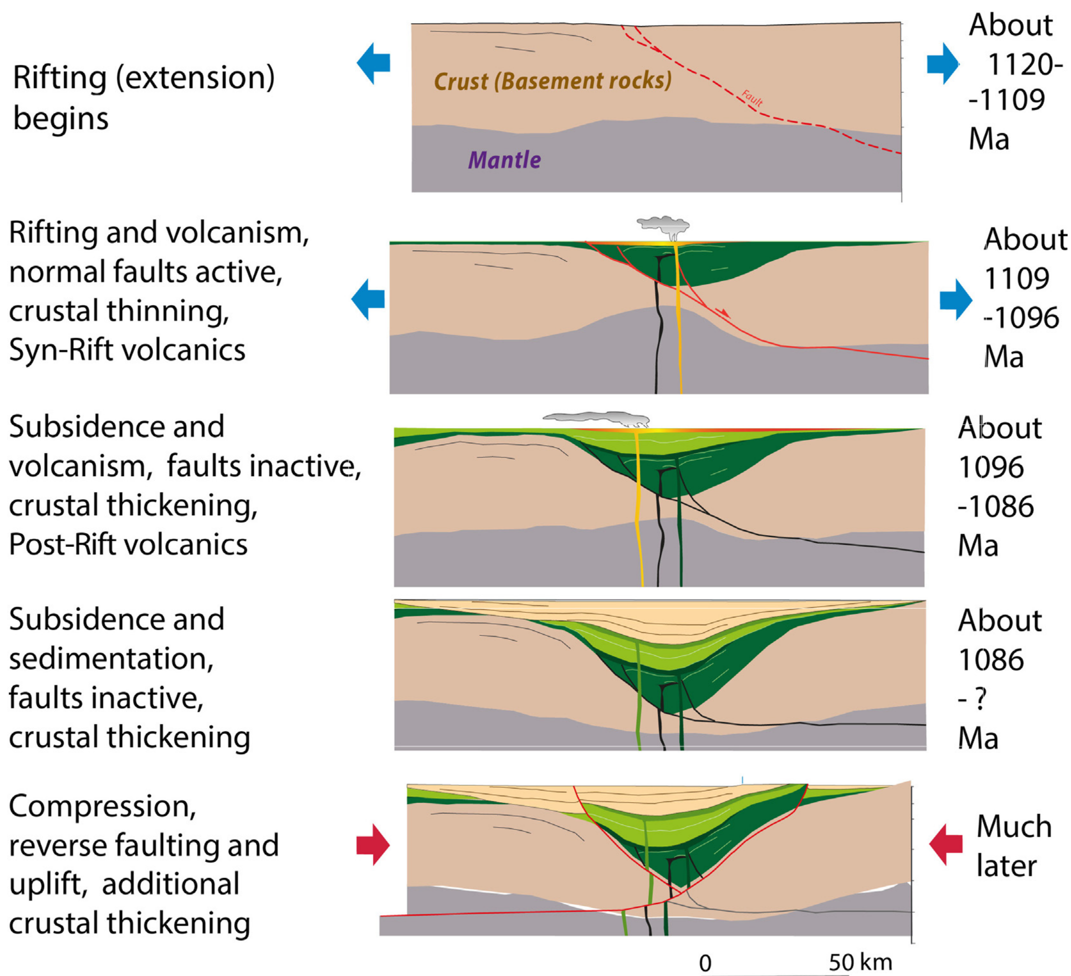

Fig. 10. Combined rift/LIP schematic model of MCR evolution based on the structural restorations. As in Fig. 9, crust (basement rocks) are shown in pink, sediment in yellow, volcanics in green, and mantle in gray. The original crust was thinned in the rifting stage, rethickened during the postrift phase, and thickened further by the later basin inversion (Stein et al., 2015).

gravity profiles indicates a magma volume of ~2×10^6 km^3 (Merino et al., 2013), well above the LIP threshold of 10^5 km^3 (Ernst, 2014). Comparing the volume and surface area of this LIP, sometimes termed the Keewenaw LIP, to other flood basalts (Fig. 8) shows that it is on average significantly thicker, because its large volume was deposited in a narrow rift rather than across a broad surface. Hence the MCR has the geometry of a rift but the igneous rock volume of a LIP.

As discussed shortly, structural reconstruction of basalts and sediments within the rift support this view. This combination resolves the paradox that the MCR's geometry and tectonic setting are consistent with "passive" rifting at a plate boundary whereas the volume and composition of the volcanic rocks are interpreted as showing that the MCR formed by "active" rifting over a mantle plume (Nicholson et al., 1997; White, 1997).

2.2. *A fault-bounded rift basin filled with volcanics and sediments*

A model for MCR evolution has been developed (Stein et al., 2015) by combining the rift/LIP concept from the gravity data with structural modeling of GLIMPCE seismic reflection lines across western Lake Superior (Green et al., 1989) that were also used in previous models (e.g., Shay and Tréhu, 1993). The profiles, such as line C (Fig. 9), show ~20 km maximum thickness of volcanics, overlain by ~5–8 km of mostly conformable sedimentary strata. From their seismic appearance and correlation with outcrops on land, the volcanics were subdivided into the younger Portage Lake series, underlain by the older pre-Portage Lake series (Hutchinson et al., 1990). Most of the basin fill is confined between two steeply inward dipping faults that flatten and converge at depth, forming a bowl-shaped depression. The upper regions of the faults show reverse offsets of stratigraphic markers, due to basin inversion (Chandler et al., 1989) long after rifting, volcanism, and subsidence ended.

The reflection data indicate a history of extension, volcanism, sedimentation, subsidence, and reverse faulting. The lower volcanic layers – primarily the pre-Portage Lake series - truncate toward the north side of the rift basin, indicating deposition during normal fault motion. However, the upper volcanic layers – primarily the Portage Lake series - and overlying Oronto postrift sediments dip from both sides and thicken toward the basin center, indicating deposition in a cooling and subsiding bowl-shaped, largely unfaulted basin. Hence the first (synrift) units were deposited during a rifting phase, whereas the second (postrift) units were deposited during thermal subsidence with no significant associated faulting after extension ended.

2.3. *A history of extension, volcanism, sedimentation, subsidence, and inversion*

This history was modeled (Fig. 10) via numerical stepwise structural restoration, working backwards from the present geometry, using Midland Valley's 2DMove software. The cross-sectional area and shape of the synrift deposits are consistent with ~20–25 km of extension on the Douglas/Ojibwa Fault. Taking the duration of synrift volcanism as

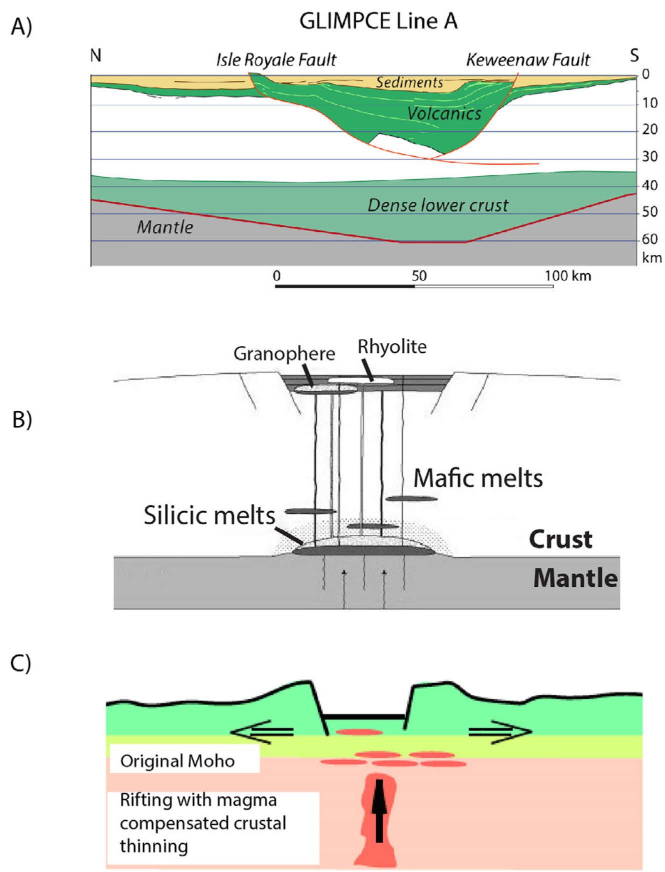

Fig. 11. Examples of underplating. A) Crustal structure model beneath Lake Superior for GLIMPCE line A showing crustal thickening (Green et al., 1989). B) Model for MCR magmatism (Vervoort et al., 2007). C) Model for magma-compensated rifting (Thybo and Artemieva, 2013).

~10 Myr (half the volcanic succession) gives an extension rate of ~2–2.5 mm/yr, a typical value for rifts. Some extension occurred before the volcanism, but its magnitude cannot be determined because sub-volcanic structures are not well imaged.

Line C shows that the Douglas/Objiwa Fault on the north side of the basin was the master fault active during rifting, whereas the Keweenaw Fault on the south side is subparallel to the base of the volcanic infill, indicating it was not a major rift-bounding normal fault during the extensional phase. In contrast, line A to the east shows that the Keweenaw fault was the master fault with 28 km of extension. This polarity reversal along a series of adjacent half-graben segments (Dickas and Mudrey, 1997) is analogous to that in the East African rift.

In this model the MCR began as a half-graben with initial largely non-volcanic extension, that was later filled by synrift and post-rift flood basalts. Some of the oldest volcanic rocks are pillow basalts emplaced in soft sand in a water-filled basin (Ojakangas and Morey, 1982), suggesting a period of rift extension and subsidence before massive flood basalt eruption.

After extension ceased, further subsidence accommodated another thick succession of flood basalts during the postrift phase. After volcanism ended, thermal subsidence continued, accompanied by postrift sedimentation. The crust was depressed and strongly flexed, deepening the Moho under the load of the dense flood basalt infill and sediment. This geometry is remarkably different from that observed beneath other continental flood basalts, where stacked and largely horizontal basalt flows without significant overlying sediment produced only minor flexure (Watts and Cox, 1989).

Extension ended about when deposition of the Portage Lake volcanics began, so about 40% of the volcanics were deposited after extension ended. Extension, volcanism, and postrift sedimentation ended long before regional compression inverted the basin by reverse fault motion (for line C, ~3 km on the Douglas/Ojibwa and ~7 km on the Keweenaw reverse faults). For line A, shortening was ~12 km on the reverse faults and at most 2 km by folding. Thus most of the basin's synclinal structure arose from postrift subsidence, not the later compression.

2.4. Crustal thickening and underplating beneath the rift basin

The crust beneath the MCR is thicker than beneath surrounding areas. In addition to tectonic thickening (Fig. 10), some thickening seems to have occurred by formation of a "rift pillow" or an "underplate" layer. The "pillow" arises because as low-density melt rises, high-density residue ("restite") ponds at the base of crust (Vervoort et al., 2007) (Fig. 11). This underplating first returned the thinned crust to its original thickness, as observed in presently-active rifts and termed "magma-compensated" rifting (Thybo and Nielsen, 2009; Thybo and Artemieva, 2013), and then thickened it further. Although the specifics of this process differ between rifts, it occurs in many cases. Additional thickening occurred when the rift was inverted (Fig. 10).

Seismic data show crustal thickening and underplating along the MCR's west arm similar to that under Lake Superior (French et al., 2009; Moidaki et al., 2013; Shen et al., 2013; Zhang et al., 2016) (Fig. 12), implying that this arm formed similarly to that in the model based on the Lake Superior data. Hence it seems reasonable to use the Lake Superior model as a general scenario, while recognizing that crustal structure along the MCR should vary depending on the amounts of extension, volcanism, compression, and underplating and the directions of rifting and compression at that portion of the MCR. For example, the increase in crustal thickness across the MCR in central Iowa, found from Florida-to-Edmonton (FLED) seismic array data, is about 10 km instead of the 20 km found with SPREE data beneath Wisconsin and Minnesota (French et al., 2009; Moidaki et al., 2013).

2.5. Small seismic velocity perturbations

The MCR appears quite differently when viewed with different types of data. The MCR's dense volcanic rocks appear clearly on gravity (Fig. 2), seismic reflection (Fig. 9), and receiver function data (Fig. 12), all of which are sensitive to density contrasts. However, surface wave tomography (Fig. 13), which is sensitive to variations in seismic velocities, finds only small contrasts between the MCR and its surroundings. At a depth of 1–3 km, these data show low velocity sedimentary rocks in the west arm but not on the east arm except in the Michigan Basin, implying thin sediments along the east arm south of the basin. At depths of 3–11 km, neither arm is evident though the Michigan Basin persists. In the middle crust, neither of the arms nor the Michigan basin stands out. In the lower crust, the west arm is slow, perhaps due to the deeper Moho from crustal thickening, but the east arm does not stand out.

Hence the surface wave data "see" the shallower low velocity sediments, gravity "sees" the underlying high-density volcanics, and reflection data "see" both. Because the volcanic rocks filling the lower portions of the rift basin are not apparent in the surface wave tomography, their shear-wave velocities do not differ significantly from those of the lower crust outside the rift basin. Although the resolution of surface wave tomography decreases with the depth, it seems unlikely that the absence of a shear-wave velocity anomaly is an artifact of low resolution. However, Shay and Tréhu's (1993) model of the GLIMPCE reflection data has P-wave velocities of the rift volcanics higher than those of the lower crust outside. These differences likely come from the V_p, V_s and density properties of the various rift and crustal rocks. Basalt, which likely accounts for most of the rift-filling volcanics, has an unusual combination of high density and low shear velocity compared to most crustal rocks, and thus an anomalously large V_p/V_s ratio

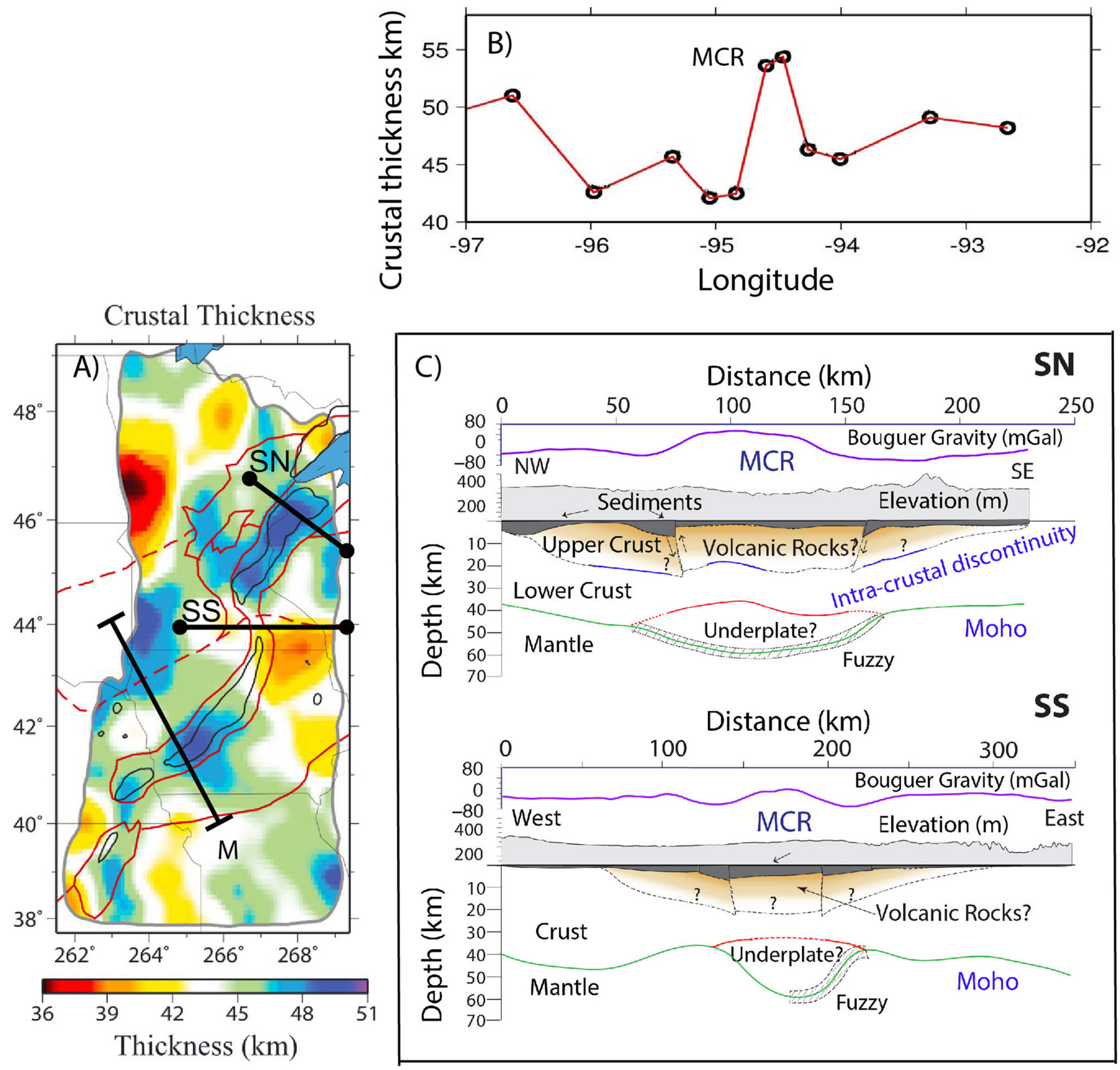

Fig. 12. Crustal thickening beneath the MCR's west arm is shown by surface wave tomography (Shen et al., 2013) (A) and receiver functions (B; Moidaki et al., 2013 and C; after Zhang et al., 2016). Profile locations are indicated in A): M denotes that in panel B) and SS and SN denote those in panel C).

(Christensen, 1996).

Another striking observation (Fig. 13e) is that the mantle below the MCR differs at most only slightly from its surroundings. P-wave teleseismic tomography shows that at 100 km depth, mantle lithosphere below the MCR is at most slightly slower than its surroundings along part of its length and indistinguishable below Lake Superior where the largest anomaly would be expected (Frederiksen et al., 2017). Similarly, shear-wave splitting data show no significant anomaly beneath the MCR, although they show a significant change across it, implying that the Superior province to the north was so thick and strong that the MCR did not break into it (Ola et al., 2015). These observations suggest that as the MCR formed, melt extraction from the mantle produced magma now filling the lower portions of the rift with shear velocity similar to and density higher than the surrounding lower crust, and left little velocity perturbation in the upper mantle (Fig. 14). Denser eclogite may also be present in the rift's lowermost crust, but gravity inversions find that mantle lithospheric density beneath the MCR is not anomalous (Levandowski et al., 2015).

Why the lithospheric mantle below the MCR shows little perturbation is unclear. By analogy to other LIPs, the melts that generated the Keewenaw LIP likely resulted from a combination of melting of the preexisting continental lithospheric mantle and plume-influenced asthenosphere (Lightfoot et al., 1993; Beccaluva et al., 2009; Trestrail et al., 2017). For these purposes, a plume can be regarded as a thermochemical anomaly without considering its geometry (Rooney, 2017). Extraction of partial melt from the mantle should have decreased the iron content of the residual mantle and resulted in the removal of garnet or spinel (depending on depth), and clinopyroxene (Ellam et al., 1992). This would be expected to occur over a range of depths from ~150 km to near the Moho, as the dominant melt source region is thought to have progressed from deeper to shallower as volcanism progressed (Nicholson et al., 1997). Some studies, based on xenolith compositions and velocities, suggest that melt depletion significantly increases mantle velocity (Jordan, 1988; Lee, 2003), whereas recent results from petrological modeling (Schutt and Lesher, 2006; Afonso and Schutt, 2012) indicate that "*melt depletion has almost no effect*" on P and S wave velocities. The absence of a major velocity anomaly below the voluminous MCR basalts argues either that the melting and modification of the preexisting lithospheric mantle had net little effect, or that the mantle beneath the MCR has been replaced since the MCR formed and moved away from where it formed.

2.6. Formation at a plate boundary

Although the MCR was traditionally assumed to have formed by isolated rifting in a plate interior (Cannon et al., 1989), it now appears

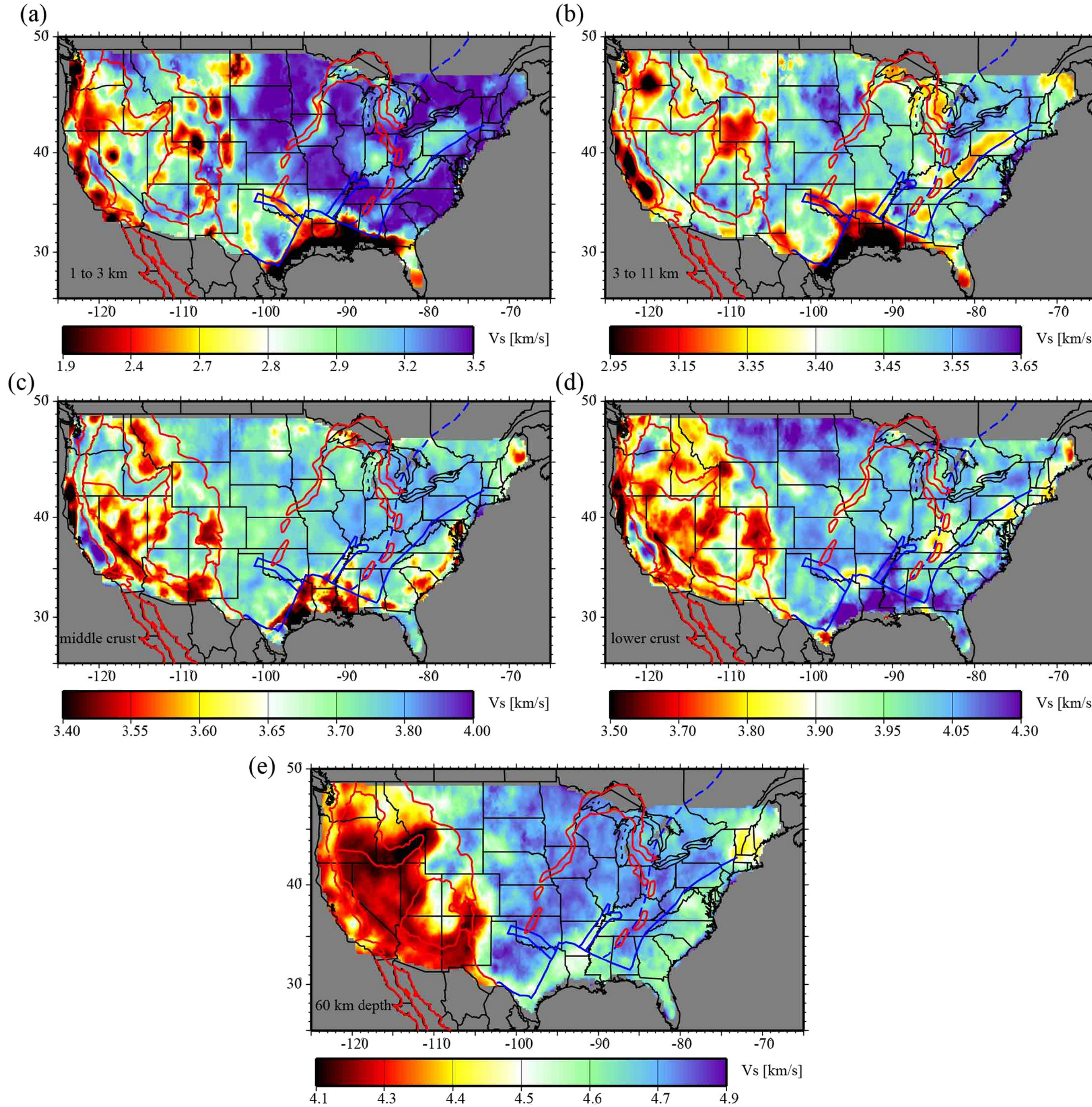

Fig. 13. S-wave velocity maps for different depths. The MCR appears as a low velocity region at 1–3 km, and has little or no clear signature at greater depths (Schmandt et al., 2015).

to have formed as part of a plate boundary reorganization. Evidence for this view comes from the change in Laurentia's absolute plate motion recorded by the MCR's volcanic rocks (Fig. 15). Such cusps in APW paths have been observed elsewhere when continents rifted apart (Gordon et al., 1984). This may have been the rifting of the Amazonia craton from Laurentia between compressional phases of the Grenville orogeny (Stein et al., 2014, 2016), although Amazonia's motion is not well constrained because of the limited paleomagnetic data available. In this view, the MCR's formation and shutdown was part of the evolution of the plate boundary between Laurentia and neighboring plates, such that the rift failed when full seafloor spreading between the major plates was established.

It seems plausible that the MCR's east and west arms acted as boundaries of a microplate within an evolving plate boundary system. In an initial model (Chase and Gilmer, 1973), both arms were boundaries between Laurentia and the microplate, such that the west arm was a spreading center and the east arm was a leaky transform. A possible consequence of this geometry is that the west arm had significantly more magma, consistent with the fact that the gravity high across along the west arm is larger than that along the east arm (Merino et al., 2013).

Another possible geometry would be analogous to today's East African Rift system, a set of microplates within the boundary zone where the major Nubian and Somalian plates diverge (Fig. 16). Present-day continental extension in the East African Rift (EAR) and seafloor spreading in the Red Sea and Gulf of Aden form a classic three-arm rift geometry as Africa splits into Nubia, Somalia, and Arabia. GPS and earthquake data show that the opening involves several microplates between the large Nubian and Somalian plates (Saria et al., 2013). If the EAR does not evolve to seafloor spreading and dies, in a billion years it

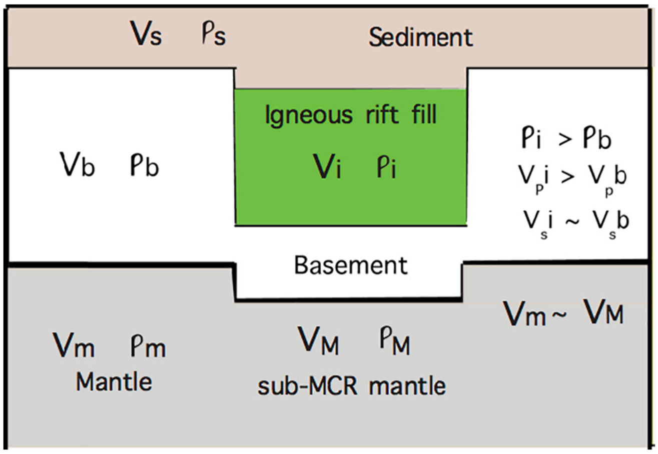

Fig. 14. Compared to the surrounding basement rock, the MCR's igneous rift fill is denser, has higher P-wave velocity, but similar or slightly lower S-wave velocity. Mantle below the MCR has P- and S-wave velocities and density similar to those of the surrounding mantle. For simplicity, underplated layer above Moho is not shown.

would appear as an isolated intracontinental failed rift similar to the MCR.

Another analogy is the West Central African Rift (WCAR) system formed as part of the Mesozoic opening of the South Atlantic. Reconstructing the fit between Africa and South America without overlaps and gaps and matching magnetic anomalies requires microplate motion with extension within continents (Moulin et al., 2010; Seton et al., 2012). These rifts failed when seafloor spreading initiated along the whole boundary between South America and Africa, illustrating that intracontinental extension can start as part of continental breakup and end when full seafloor spreading is established.

Modeling microplate evolution is straightforward for presently active systems, where relative motions along the boundaries can be found using GPS, marine magnetic, bathymetric, or earthquake slip vector data (e.g., Engeln and Stein, 1984; Engeln et al., 1988; Saria et al., 2013) and inverted to find Euler vectors. Typically the Euler poles for the microplate relative to each of the major plates are nearby, because relative velocities vary rapidly in direction and rate along the microplate's boundaries and small changes in Euler vectors correspond to the system's relatively rapid evolution. The difference between the two microplate Euler vectors equals the Euler vector for the two major plates, because the motion across the microplate sums to that between the major plates.

Fig. 17 shows one possible model for the MCR, assuming an "Illinois" microplate between the major Amazonia and Laurentia plates. Some constraints for Euler vectors can be taken from the geometry of the MCR arms. The west arm has offsets that can be treated as transforms (Chase and Gilmer, 1973), and spreading can be assumed to be approximately orthogonal to the east arm. Under these assumptions, spreading in the Lake Superior area would have been oblique. The estimated extension rate between Illinois and Laurentia is from the structural reconstructions (Fig. 10), and that along the east arm is assumed to be 50% less. Chase and Gilmer (1973) inferred spreading rates from the width of the gravity anomaly, which may be biased because the anomaly also reflects the geometry of the later compression.

The model shown, though not well constrained, seems generally plausible. It illustrates average motions over 10 million years during which the MCR was extending and had extensive volcanism as Amazonia was starting to break slowly away from Laurentia. We lack information about the major plate geometry - i.e. where to draw boundaries - and information about the southern microplate boundary is lacking due to subsequent collisions and the latest Precambrian/Cambrian rifting event (Thomas et al., 2012). For simplicity, we assume that Amazonia did not extend far south of the microplate.

3. Comparing the MCR to other rifts

We suggest that the sequence of rifting, volcanism, sedimentation, subsidence and later compression (Fig. 10) that gave rise to the MCR today gives insight into other rifts. In particular, the MCR came close to seafloor spreading before it failed, and thus records a late stage of rifting in which a large volume of magma accumulated both during extension and after extension stopped.

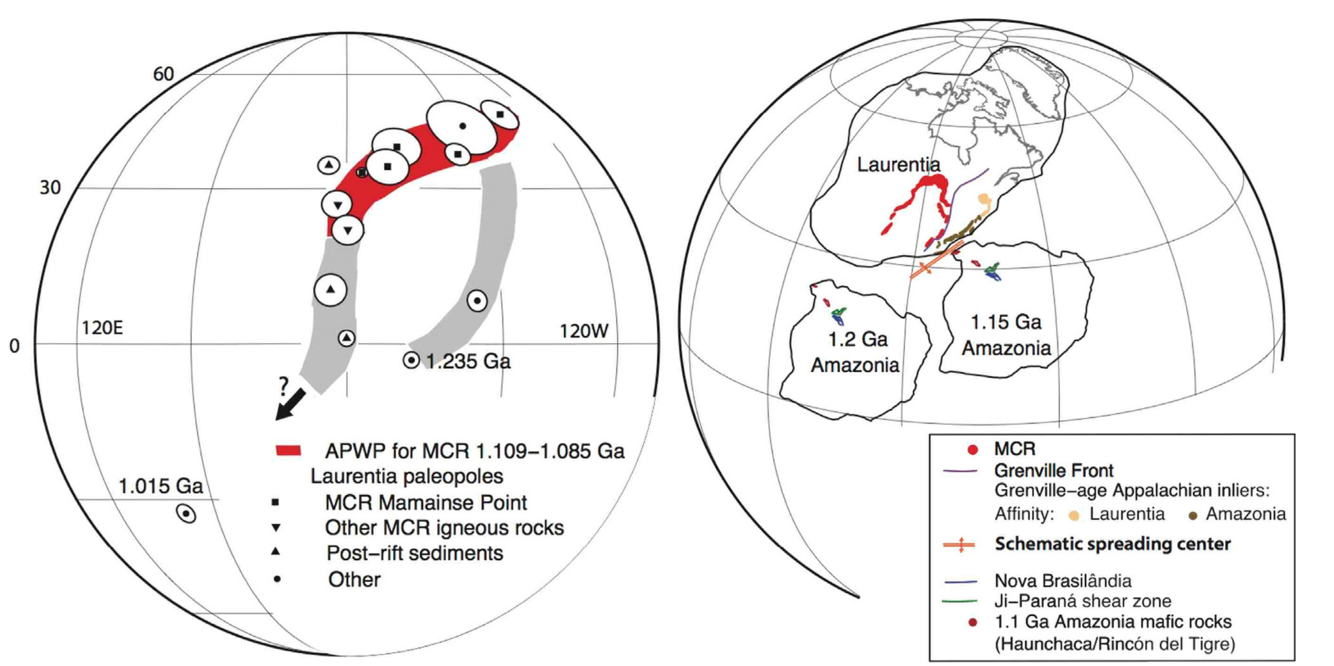

Fig. 15. (Left) Apparent polar wander path for Laurentia, showing change in motion approximately at onset of MCR volcanism (1.109 Ga). MCR volcanism interval is shown in red. (Right) Reconstruction of plate positions before Laurentia-Amazonia separation, schematic spreading center geometry, and relevant features (Stein et al., 2014).

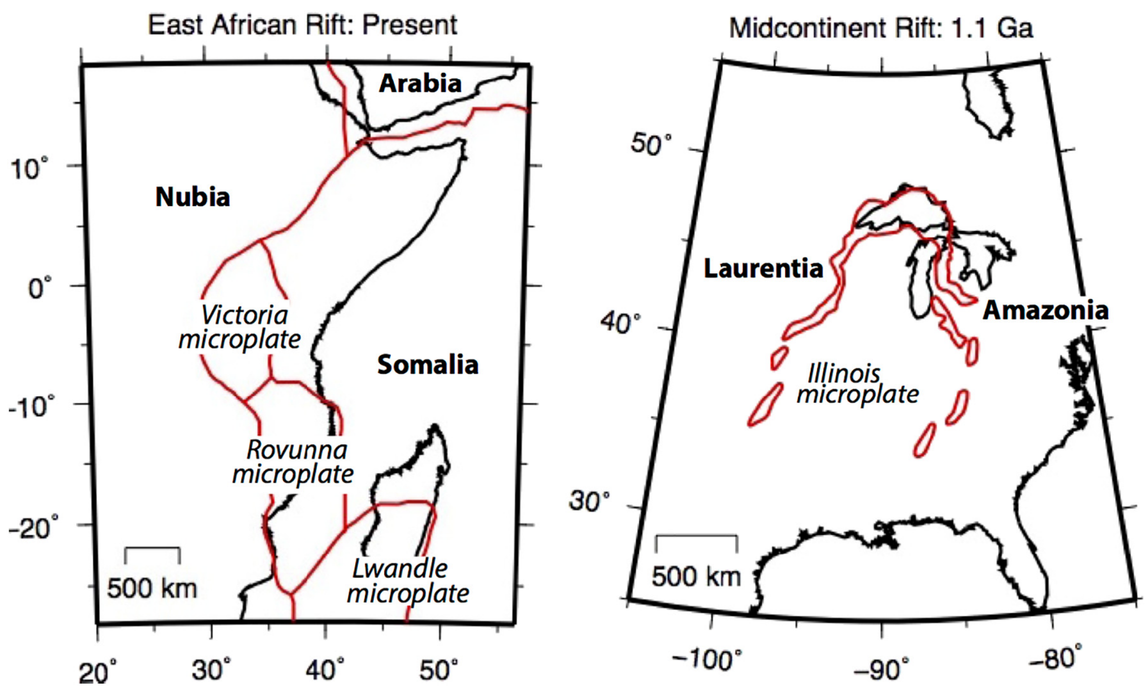

Fig. 16. Left: Geometry of microplates along the East African Rift system within the boundary zone where the major Nubian and Somalian plates diverge (after Saria et al., 2013). Right: A possible MCR geometry with an "Illinois" microplate between the diverging Amazonia and Laurentia major plates.

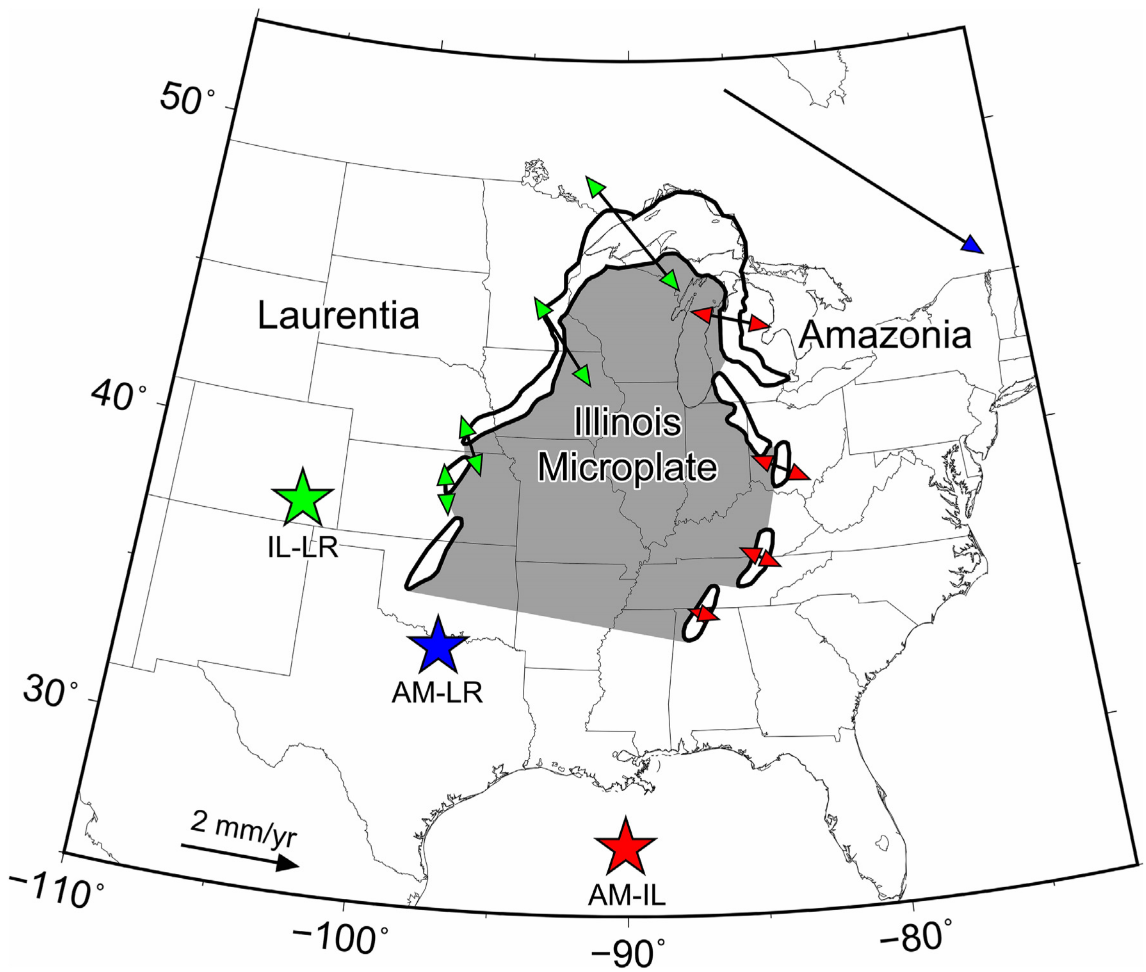

Fig. 17. Kinematic model of an "Illinois" (IL) microplate for which the MCR arms are plate boundaries. Euler poles are shown by stars, with first plate listed rotating clockwise with respect to the second. Double-headed arrows show relative motion across MCR arms, single-headed arrow shows Amazonia (AM) motion relative to Laurentia (LR). Rate scale shown by 2 mm/yr arrow.

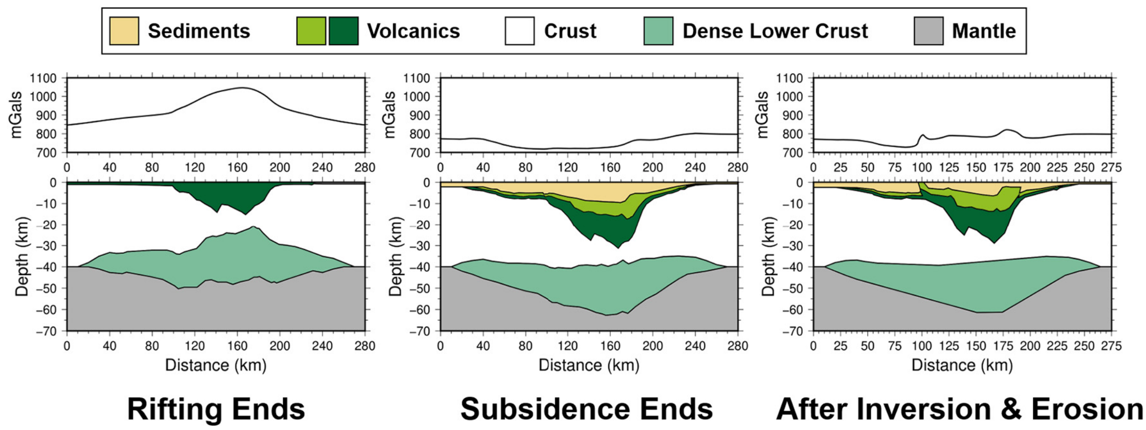

Fig. 18. Models of the gravity anomaly expected at various stages of the MCR's evolution.

No two rift systems are the same, and there are important differences in structure and evolution even among different segments of a given rift system. Still, we think it useful to view many rifts as following a generally similar sequence of evolution, shown schematically in Fig. 7. Hence we can compare how we think the MCR looked at different stages of its evolution to other rifts that are presently at similar stages. Although none are exactly comparable to the MCR, the comparison provides useful insights.

In this sprit, we use the structural model in Fig. 10, derived for the Lake Superior region, to infer the gravity anomaly that would have been expected at different stages in the MCR's evolution (Fig. 18). In the early rifting stages, the dense volcanics near the surface would have caused a large positive gravity anomaly. Subsequent deposition of low-density sediments and associated subsidence that depressed the volcanics would have caused a gravity low. Eventually, inversion of the rift and erosion that brought the volcanics closer to the surface would have caused the gravity high observed today. This sequence shows why the MCR is characterized by a prominent positive Bouguer anomaly (Fig. 2) due to the thick high-density volcanics filling it that have been uplifted by the rift basin inversion. Fig. 18 (right) shows why the Lake Superior region has a relative gravity low above the thick central rift sediments and a relative high above the reverse faults. This central low does not appear for the arms because the remaining rift sediments are much thinner.

In contrast, presently extending continental rifts, such as the Rio Grande rift, have much less volcanic fill than the MCR (Fig. 19). As a result, they have negative gravity anomalies because they are largely filled with low-density sediment, whose effects overwhelm that of the higher density mantle at shallow depth due to the crustal thinning associated with the extension.

Fig. 20 shows two other examples. The presently-active East African rift looks much like what we envision for the MCR during its rifting phase: two arms, both actively extending, with gravity lows over the arms and crustal thinning beneath them. Presumably if extension ceased then after ~100 Myr cooling the low density in the mantle would no longer be visible. The Southern Oklahoma Aulacogen, a failed rift that opened in the Early Cambrian as part of Rodinia's breakup and was inverted in the late Paleozoic, shows a gravity high due to the igneous rocks filling the rift, and thickened crust. This looks like a mini-MCR, with similar general features although the igneous fill is smaller and shallower than the MCR's.

Failed continental rifts differ in their level of modern seismic activity. Some failed continental rifts present weak places within the continent likely to take up strains in response to new stress regimes. For example, though the Reelfoot Rift and Wabash Valley Faults initially formed as rifts, they are now locations of moderate amounts of compressional intraplate seismicity due to the overall compressional stresses currently within the mid-continent. This is not the case with the MCR, however, which has anomalously low levels of intraplate seismicity. Analysis of two years of SPREE data found only 12 intraplate earthquakes within Wisconsin, Minnesota, Upper Michigan, and Ontario, with none of these larger than magnitude 3, and only one occurring within the MCR itself (beneath western Lake Superior) (Bartz et al., 2014). Unless this low seismicity simply reflects the short time

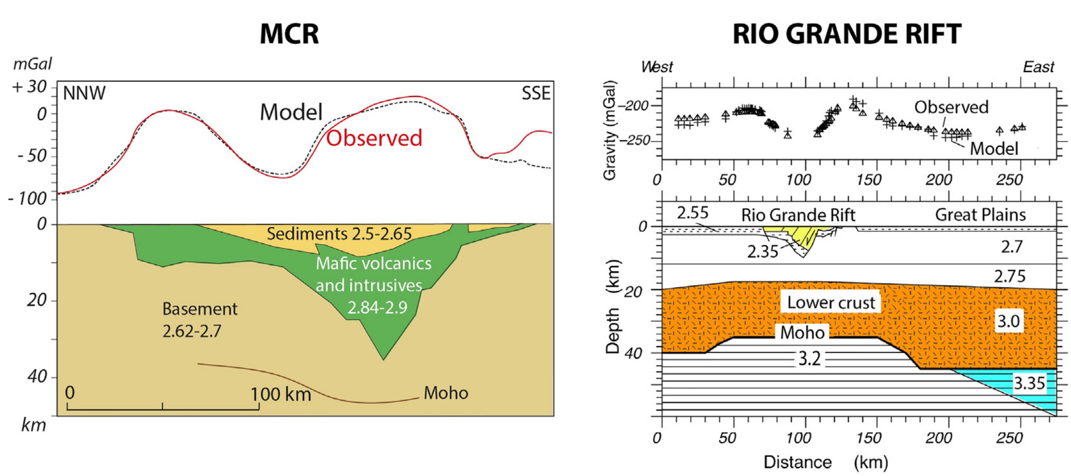

Fig. 19. Left) Bouguer gravity data and model across the MCR, showing positive anomalies due to high-density volcanics. Red line is observed gravity; black dots are calculated (Thomas and Teskey, 1994). Right) Bouguer gravity data and integrated geophysical model across the Rio Grande rift, showing negative anomalies due to low-density sediments (Grauch et al., 1999, 2015).

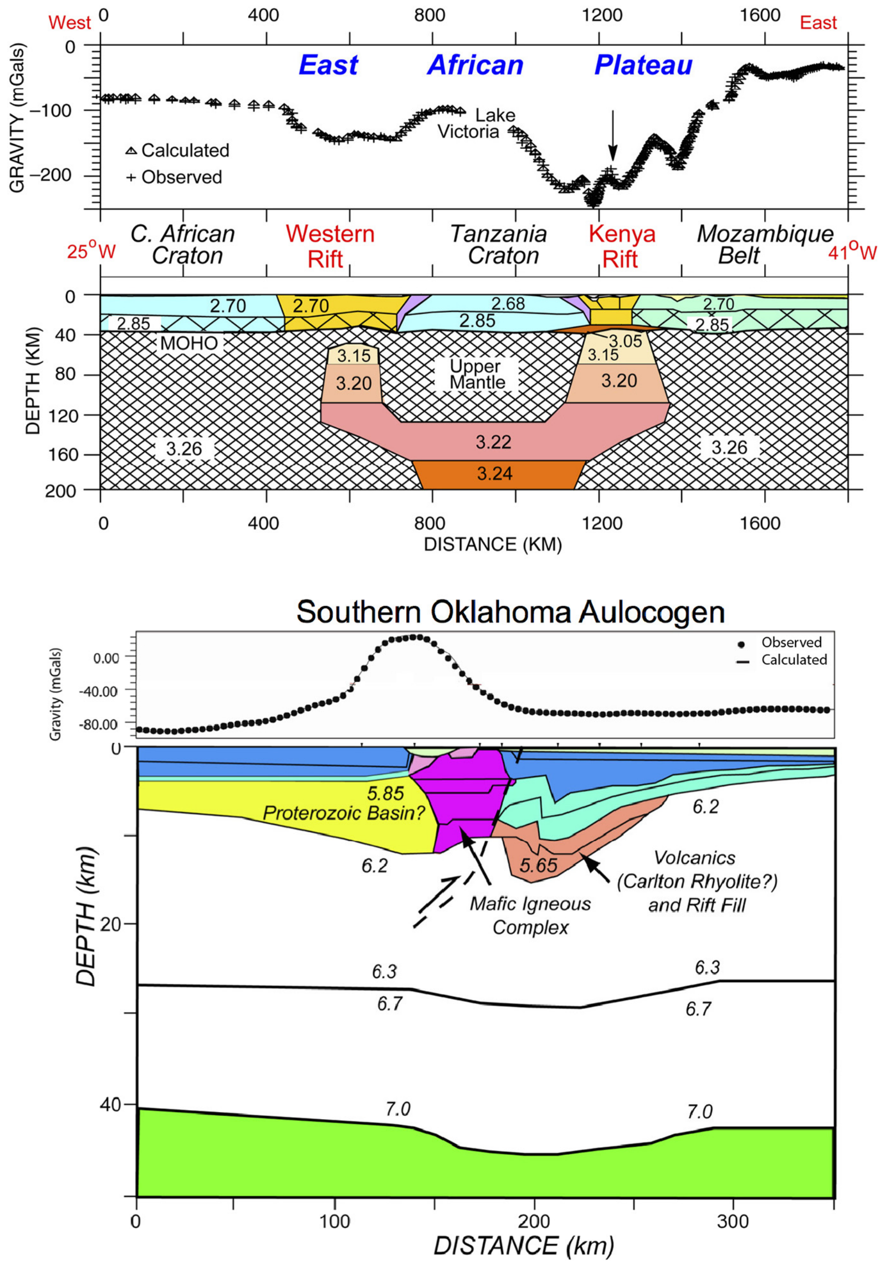

Fig. 20. Cross-sections of the presently-active East African rift (top) (Simiyu and Keller, 1997), a good analogy to the MCR's early stages, and (bottom) the failed and massively inverted Southern Oklahoma Aulacogen (Hanson et al., 2013), which is similar to the MCR today.

sampled (Liu and Stein, 2016), the MCR here seems to be too thick and strong to allow present-day reactivation of the faults within it. However, southern extensions of the MCR across Kansas and Oklahoma have experienced seismicity and Phanerozoic deformation (Luza and Lawson, 1981; Burberry et al., 2015).

4. Comparing the MCR to passive continental margins

The MCR has interesting implications for volcanic passive continental margins. Volcanic margins, the most common passive margins, arise where continental breakup is associated with the eruption of flood basalts during prerift and/or synrift stages of continental separation, in which large-scale melting gives rise to thick igneous crust (Menzies et al., 2002; Geoffroy, 2005; Geoffroy et al., 2015). These margins are characterized by seaward dipping reflectors, volcanic rocks yielding magnetic anomalies landward of the oldest spreading anomalies, thinned continental crust, and a high-velocity lower crustal (HVLC) body (Fig. 21).

The MCR gives insight into how rifting and volcanism interacted in the early phase of volcanic margin formation, a record that is lost if a rift evolves into seafloor spreading. The MCR came close to seafloor spreading before it failed, and thus records a late stage of rifting in which a large volume of magma accumulated both during extension and after extension stopped. The analogy with volcanic margins is not perfect, in that the upper portion of the MCR volcanics were deposited after extension ended. However, because the MCR has many of the features of passive margins, it is useful to view the MCR as a preserved piece of what might have evolved to a volcanic margin had the MCR not failed, but instead split and started seafloor spreading (Fig. 22). Many of the key features seen at passive margins would have formed this way, though they would be modified as seafloor spreading developed. High-quality seismic reflection data show that packages of SDRs form over long periods of time during extension (Blaich et al., 2011). Additional SDRs are deposited as seafloor spreading starts (Koopmann et al., 2014)

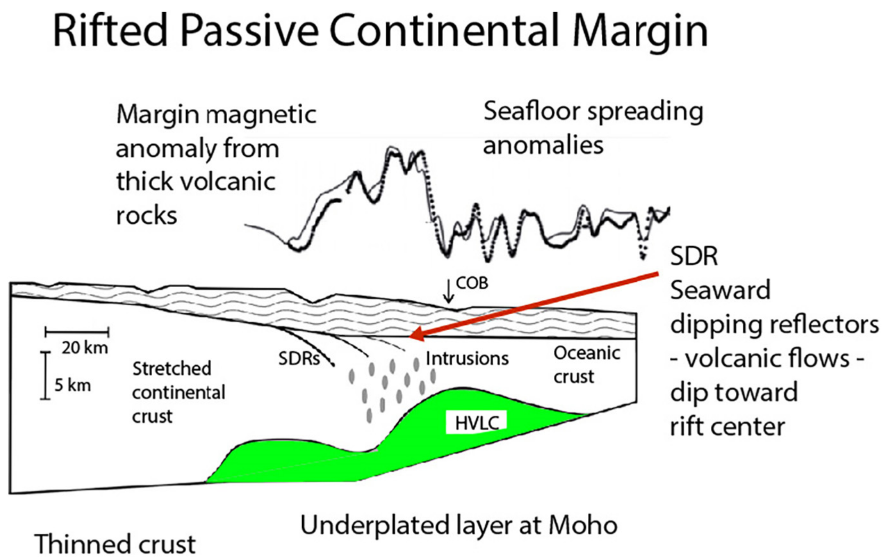

Fig. 21. Schematic cross section through a rifted volcanic passive margin, showing characteristic tectonic elements (Schnabel et al., 2008; Koopmann et al., 2014). COB denotes continent-ocean boundary, SDR denotes seaward dipping reflectors, and HVLC denotes high-velocity lower crustal bodies.

and the SDRs acquire concave-down curvature due to flexure (Buck, 2017).

Continued successful half-graben rifting would have yielded asymmetric passive margins with the key features observed at present-day margins. These are often observed on opposite sides of conjugate margins, where one side is wider than the other (Fig. 23).

Key dimensions, including magma volumes, are comparable between the MCR and present-day passive margins. Specifically, the cross-sectional volume of the MCR volcanics beneath Lake Superior is similar to that of the combined conjugate sides of the volcanic margin shown in Fig. 23. Hence one can view the MCR as a passive margin you can walk around on (Fig. 24). The fact that the MCR shows many features of a rifted volcanic margin suggests that it came close to continental breakup before it failed, and illustrates how many passive margin features form prior to breakup.

5. Discussion

The MCR's history illustrates that many rifts can be viewed as following a generally similar evolutionary sequence, within which a complex combination of factors control the variability of structures within and between rifts.

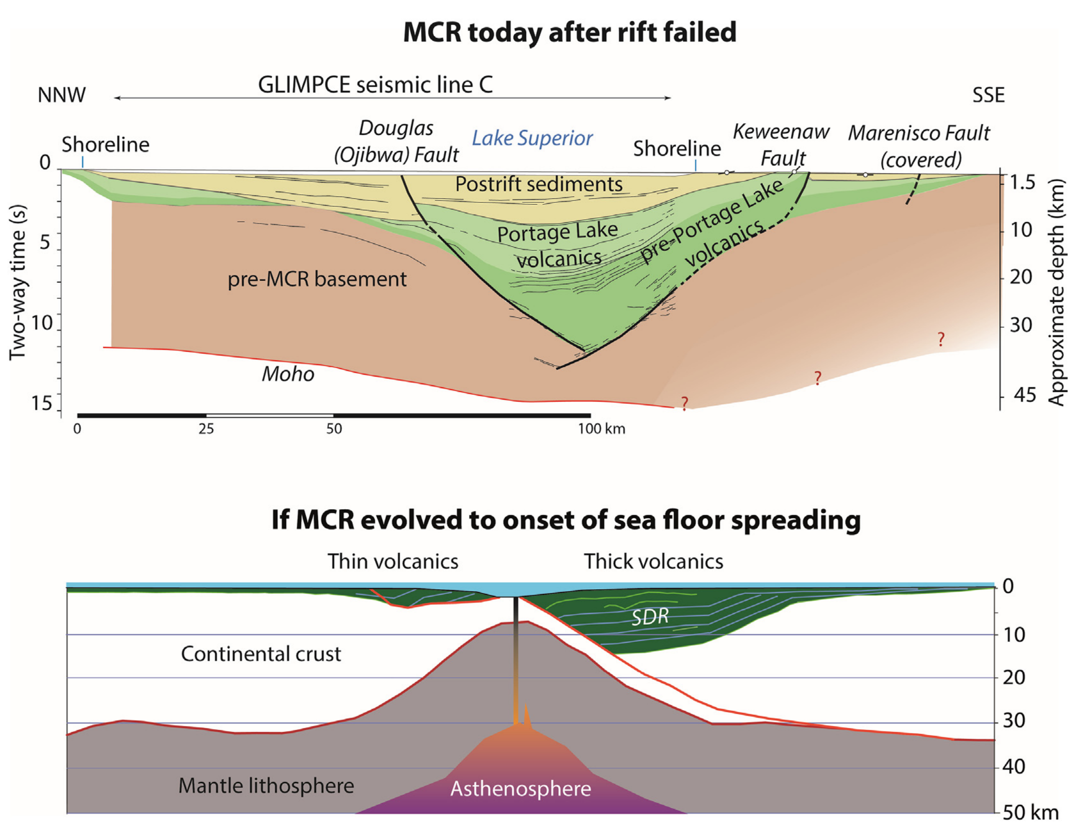

Fig. 22. Comparison of present MCR structure (top) to conceptual model for the MCR at the onset of seafloor spreading if it had not failed (bottom).

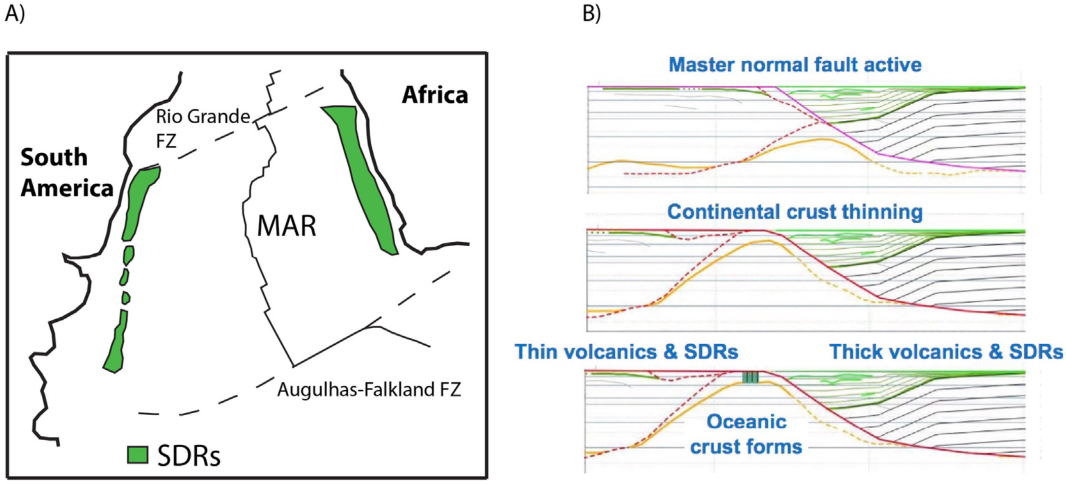

Fig. 23. Left: Conjugate passive margins in the South Atlantic. The zone of SDRs and volcanics (green) is wider on the African side (after Blaich et al., 2011). Right: Geometric model of how the MCR could have yielded an asymmetric passive margin after continued rifting. Grid of lines on right-hand side represents model strain markers, not crustal structure.

For example, the gravity high across the MCR's west arm is larger and is bounded by pronounced lows, whereas that across the east arm is smaller and lacks the sharp bounding lows (Fig. 25). A variety of explanations can be offered, because the present structure reflects the combined effects of a sequence of rifting, volcanism, sedimentation, subsidence, compression (Fig. 10), and any later effects. Differences in any of these would cause differences in the final structure and the resulting gravity anomaly (Fig. 18). One possible cause of the different heights is more magma in the west arm (Merino et al., 2013). Another is that the east arm was subjected to less inversion than the west arm, yielding a smaller gravity high but without the bounding lows. Intuitively this seems implausible, because any Grenville compression should have been stronger on the east arm. A third alternative is that the east arm was more strongly inverted than the west arm, such that some of the igneous rock was eroded.

Differences also appear between the MCR and two younger failed rifts in North America, the massively-inverted Southern Oklahoma Aulacogen and slightly-inverted Reelfoot Rift. The rifts appear differently in gravity (Fig. 2), magnetic (Fig. 3), and seismic (Fig. 13) data. In our view, they likely followed grossly similar evolutionary paths, but with differences in the extent of each stage.

An analogy might be the way differences between the terrestrial planets - Earth, Mars, Venus and the moon (a planet for these purposes) - arise. Kaula (1975) proposed that these follow a generally similar sequence of phases including their formation, early convection and core formation, plate tectonics, terminal volcanism, and quiescence. The extent to which these operated controls each planet's state. Earth is in its middle age, characterized by active plate tectonics, whereas the smaller and thus colder moon is quiescent. Although this idea does not explain all of the differences between planets, notably that Earth is the

Fig. 24. South-dipping basalt flows in Isle Royale National Park, Michigan, on the north side of the MCR rift basin (Fig. 9), are analogous to seaward dipping reflectors at a passive continental margin.
Photo by Seth Stein

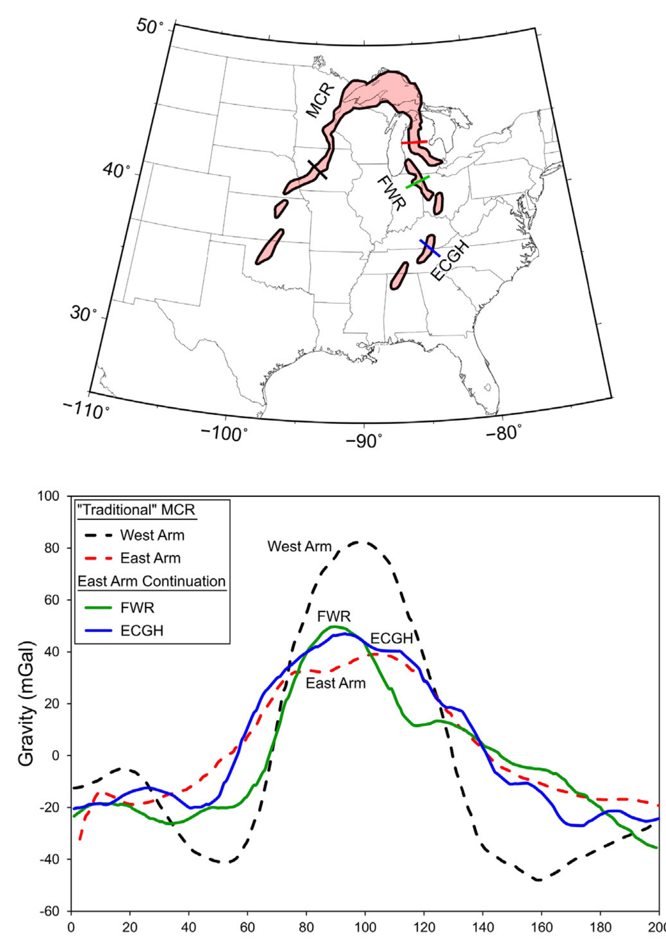

Fig. 25. Gravity profiles across the east and west arms of the Midcontinent Rift (Stein et al., 2018).

only one with plate tectonics at present, it is a useful starting point.

Our view that the MCR illustrates aspects of the process by which continental rifting gives rise to volcanic passive margins leads to the long-standing and unresolved question of the role of mantle plumes in continental breakup. The question can be posed as: what controls which rifts acquire such large volumes of magmatic rocks? It has long been proposed that excess temperatures associated with hotspots/plumes are required to rift continents apart, given the volumes of igneous rocks at most passive margins (Burke and Whiteman, 1973; Morgan, 1981, 1983; White and McKenzie, 1989; Richards et al., 1989). High temperatures are also inferred for the MCR from petrologic and geochemical data (Nicholson et al., 1997; White, 1997). However, invoking plumes for all LIPs and rifted margins has been questioned and alternatives have been proposed (King and Anderson, 1995; King, 2007; Foulger, 2011). van Wijk et al.'s (2001, 2004) analyses favor generation of volcanic margins by decompression melting alone without the aid of mantle plumes. Franke (2012) finds that the rifting and spreading history of the South Atlantic, a classic volcanic margin (Fig. 21), cannot be reconciled with a mantle plume model. How to generate long linear rifts from a plume remains unclear and under investigation (Koptev et al., 2017; Beniest et al., 2017). As our results give no direct insight into these issues, we leave this topic for future studies.

Acknowledgements

This work was supported by NSF grants EAR-1550108, EAR-1148088, EAR-1549764, EAR-0952345, and EAR-0952154. We thank Fan-Chi Lin for providing the velocity maps in Fig. 13, and the community of MCR enthusiasts for helpful discussions. We have benefitted from helpful reviews by Jolante van Wijk and Stephen Marshak.

References

Adams, D.C., Keller, G.R., 1994. Possible extension of the Midcontinent Rift in West Texas and eastern New Mexico. Can. J. Earth Sci. 31 (4), 709–720.
Adams, D.C., Keller, G.R., 1996. Precambrian basement geology of the Permian Basin region of West Texas and eastern New Mexico: a geophysical perspective. AAPG Bull. 80 (3), 410–431.
Afonso, J.C., Schutt, D.L., 2012. The effects of polybaric partial melting on density and seismic velocities of mantle restites. Lithos 134, 289–303.
Al-Eqabi, G., Wiens, D., Wysession, M.E., Shen, W., van der Lee, S., Revenaugh, J., Frederiksen, A.W., Darbyshire, F.A., Stein, S.A., Jurdy, D.M., Wolin, E., Bollmann, T.A., 2015. Rayleigh Wave tomography of Mid-Continent Rift using earthquake and ambient noise data. In: AGU Fall Meeting Abstract.
Allen, P.A., Armitage, J.J., 2012. Cratonic Basins. John Wiley Sons, Ltd.
Baranoski, M.T., Dean, S.L., Wicks, J.L., Brown, V.M., 2009. Unconformity-bounded seismic reflection sequences define Grenville-age rift system and foreland basins beneath the Phanerozoic in Ohio. Geosphere 5, 140–151.
Bartholomew, M.J., Hatcher, R.D., 2010. The Grenville orogenic cycle of southern Laurentia: unraveling sutures, rifts, and shear zones as potential piercing points for Amazonia. J. S. Am. Earth Sci. 29, 4–20.
Bartz, D., Wysession, M.E., Wiens, D.A., Aleqabi, G.I., Shore, P., van der Lee, S., Jurdy, D., Stein, S., Revenaugh, J., Wolin, E., Bollmann, T.A., Frederiksen, A.W., Darbyshire, F.A., 2014. Assessing microseismicity of the Northern Mid-Continent Rift Zone and surrounding regions. In: AGU Fall Meeting Abstract.
Beccaluva, L., Bianchini, G., Natali, C., Siena, F., 2009. Continental flood basalts and mantle plumes: a case study of the Northern Ethiopian Plateau. J. Petrol. 50, 1377–1403. https://doi.org/10.1093/petrology/egp024.
Beniest, A., Koptev, A., Leroy, S., Sassi, W., Guichet, X., 2017. Two-branch break-up systems by a single mantle plume: insights from numerical modeling. Geophys. Res. Lett. 44 (19), 9589–9597.
Blaich, O.A., Faleide, J.I., Tsikalas, F., 2011. Crustal breakup and continent-ocean transition at South Atlantic conjugate margins. J. Geophys. Res. Solid Earth 16 (B1) (doi:0148-0227/11/2010JB007686).
Braile, L.W., Hinze, W.J., Keller, G.R., Lidiak, E.G., Sexton, J.L., 1986. Tectonic development of the New Madrid rift complex, Mississippi embayment, North America. Tectonophysics 131 (1–2), 1–21.
Bright, R.M., Amato, J.M., Denyszyn, S.W., Ernst, R.E., 2014. U-Pb geochronology of 1.1 Ga diabase in the southwestern United States: testing models for the origin of a post-Grenville large igneous province. Lithosphere 6, 135–156.
Buck, W.R., 2017. The role of magmatic loads and rift jumps in generating seaward dipping reflectors on volcanic rifted margins. Earth Planet. Sci. Lett. 466, 62–69.
Buening, J.D., 2013. Integrated Geophysical and Geological Study of the Relationships Between the Grenville Orogen and Mid-continent Rift System (M.S. thesis). University of Oklahoma, Norman (83 pp.).
Burberry, C.M., Joeckel, R.M., Korus, J.T., 2015. Post-Mississippian tectonics of the Nemaha Tectonic Zone and Mid-Continent Rift System, SE Nebraska and N Kansas. Mt. Geol. 52 (4), 47–73.
Burke, K., Whiteman, A.J., 1973. Uplift, rifting and break-up of Africa. In: Tarling, D.H., Runcorn, S.K. (Eds.), Implications of Continental Drift to the Earth Sciences. Academic Press, London, pp. 735–755.
Cannon, W.F., et al., 1989. The North American Midcontinent rift beneath Lake Superior from GLIMPCE seismic reflection profiling. Tectonics 8, 305–332.
Chandler, V.W., McSwiggen, P.L., Morey, G.B., Hinze, W.J., Anderson, R.R., 1989. Interpretation of seismic-reflection, gravity, and magnetic data across middle Proterozoic mid-continent rift system, northwestern Wisconsin, eastern Minnesota, and Central Iowa. Am. Assoc. Pet. Geol. Bull. 73, 261–275.
Chase, C.G., Gilmer, T.H., 1973. Precambrian plate tectonics – midcontinent gravity high. Earth Planet. Sci. Lett. 21 (1), 70–78.
Christensen, N.I., 1996. Poisson's ratio and crustal seismology. J. Geophys. Res. 101, 3139–3156.
Craddock, J.P., Konstantinou, A., Vervoort, J.D., Wirth, K.R., Davidson, C., Finley-Blasi, L., Juda, N.A., Walker, E., 2013. Detrital zircon provenance of the Mesoproterozoic Midcontinent Rift, Lake Superior Region, USA. J. Geol. 121, 57–73.
Dalziel, I.W., 1991. Pacific margins of Laurentia and East Antarctica-Australia as a conjugate rift pair: evidence and implications for an Eocambrian supercontinent. Geology 19 (6), 598–601.
Dickas, A.B., Mudrey, M.G., 1997. Segmented structure of the middle Proterozoic Midcontinent Rift System, North America. In: Ojakangas, R.W., Dickas, A.B., Green, J.C. (Eds.), Middle Proterozoic to Cambrian Rifting, Central North America. 312. Geological Society of America Special Paper, Boulder, Colorado, pp. 37–46.
Dickas, A.B., Mudrey, M.G., Ojakangas, R.W., Shrake, D.L., 1992. A possible southeastern extension of the Midcontinent Rift System in Ohio. Tectonics 11, 1406–1414.
Drahovzal, J.A., Harris, D.C., Wickstrom, L.H., Walker, D., Baranoski, M.T., Keith, B., Furer, L.C., 1992. The East Continent Rift Basin: A New Discovery. Ohio Division of Geological Survey.
Drenth, B.J., Anderson, R.R., Schulz, K.J., Feinberg, J.M., Chandler, V.W., Cannon, W.F., 2015. What lies beneath: geophysical mapping of a concealed Precambrian intrusive complex along the Iowa–Minnesota border. Can. J. Earth Sci. 52 (5), 279–293.
Ellam, R.M., Carlson, R.W., Shirey, S.B., 1992. Evidence from Re–Os isotopes for plume–lithosphere mixing in Karoo flood basalt genesis. Nature 359, 718–721. https://doi.org/10.1038/359718a0.
Engeln, J.F., Stein, S., 1984. Tectonics of the Easter plate. Earth Planet. Sci. Lett. 68 (2), 259–270.
Engeln, J.F., Stein, S., Werner, J., Gordon, R.G., 1988. Microplate and shear zone models for oceanic spreading center reorganizations. J. Geophys. Res. Solid Earth 93 (B4), 2839–2856.
Ernst, R.E., 2014. Large Igneous Provinces. Cambridge University Press (661 pp.).
Foulger, G.R., 2011. Plates vs Plumes, A Geological Controversy. Wiley (364 pp.).
Franke, D., 2012. Rifting, lithosphere breakup and volcanism: comparison of magma-poor and volcanic rifted margins. Mar. Pet. Geol. https://doi.org/10.1016/j.marpetgeo.2012.11.003.
Frederiksen, A., Bollmann, T.A., van der Lee, S., Wolin, E., Revenaugh, J., Wiens, D., Darbyshire, F.A., Aleqabi, G.I., Wysession, M.E., Stein, S., Jurdy, D.M., 2017. One billion yearold Midcontinent Rift leaves virtually no clues in the mantle. In: AGU Fall Meeting Abstract.
French, S.W., Fischer, K.M., Syracuse, E.M., Wysession, M.E., 2009. Crustal structure beneath the Florida-to-Edmonton broadband seismometer array. Geophys. Res. Lett. 36, L08309. https://doi.org/10.1029/2008GL036331.
Frizon de Lamotte, D., Fourdan, B., Leleu, S., Leparmentier, F., Clarens, P., 2015. Style of rifting and the stages of Pangea breakup. Tectonics 34 (5), 1009–1029.
Geoffroy, L., 2005. Volcanic passive margins. Compt. Rendus Geosci. 337 (16), 1395–1408.
Geoffroy, L., Burov, E.B., Werner, P., 2015. Volcanic passive margins: another way to break up continents. Sci. Rep. 5, 14828.
Gordon, R.G., Cox, A., O'Hare, S., 1984. Paleomagnetic Euler poles and the apparent polar wander and absolute motion of North America since the Carboniferous. Tectonics 3, 499–537.
Grauch, V.J.S., Gillespie, C.L., Keller, G.R., 1999. Discussion of new gravity maps for the Albuquerque area. In: New Mexico Geological Society Guidebook, 50th Field Conference.
Green, J.C., 1983. Geologic and geochemical evidence for the nature and development of the Middle Proterozoic (Keweenawan) Midcontinent Rift of North America. Tectonophysics 94, 413–437. https://doi.org/10.1016/0040-1951(83)90027-6.
Green, A.G., Cannon, W.F., Milkereit, B., Hutchinson, D.R., Davidson, A., Behrendt, J.C., Spencer, C., Lee, M.W., Morel-á-LáHuissier, P., Agena, W.F., 1989. A "GLIMPCE" of the deep crust beneath the Great Lakes. In: Mereu, R.F., Mueller, S., Fountain, D.M. (Eds.), Properties and Processes of Earth's Lower Crust. Geophysical Monograph Series 51. American Geophysical Union, pp. 65–80.
Hanson, R.E., Puckett, R.E., Keller, G.R., Brueseke, M.E., Bulen, C.L., Mertzman, S.A., Finegan, S.A., McCleery, D.A., 2013. Intraplate magmatism related to opening of the southern Iapetus Ocean, Cambrian Wichita igneous province in the Southern Oklahoma rift zone. Lithos 174, 57–70.
Hinze, W.J., Allen, D.J., Braile, L.W., Mariano, J., 1997. The Midcontinent Rift System, a major Proterozoic continental rift. In: Ojakangas, R.W., Dickas, A.B., Green, J.C. (Eds.), Middle Proterozoic to Cambrian Rifting, Central North America. 312. Geological Society of America Special Paper, Boulder, Colorado, pp. 7–34.
Hoffman, P.F., 1988. United plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annu. Rev. Earth Planet. Sci. 16 (1), 543–603.
Hoffman, P.F., 1991. Did the breakout of Laurentia turn Gondwanaland inside-out? Science 252 (5011), 1409–1412.

Hutchinson, D.R., White, R.S., Cannon, W.F., Schulz, K.J., 1990. Keweenaw hot spot, geophysical evidence for a 1.1 Ga mantle plume beneath the Midcontinent Rift System. J. Geophys. Res. 95, 10,869–10,884. https://doi.org/10.1029/JB095iB07p10869.
Jordan, T.H., 1988. Structure and formation of continental tectosphere. J. Petrol. 1, 11–37.
Kaula, W.M., 1975. The seven ages of a planet. Icarus 26, 1–15.
Keller, G.R., Bland, A.E., Greenberg, J.K., 1982. Evidence for a major Late Precambrian tectonic event (rifting?), in the eastern Midcontinent region, United States. Tectonics 1, 213–223.
King, S.D., 2007. Hotspots and edge-driven convection. Geology 35, 223–226.
King, S.D., Anderson, D.L., 1995. An alternative mechanism of flood basalt formation. Earth Planet. Sci. Lett. 136, 269–279.
King, E.R., Zietz, I., 1971. Aeromagnetic study of the midcontinent gravity high of Central United States. Geol. Soc. Am. Bull. 82, 2187–2208.
Koopmann, H., Schreckenberger, B., Franke, D., Becker, K., Schnabel, M., 2014. The late rifting phase and continental break-up of the southern South Atlantic: the mode and timing of volcanic rifting and formation of earliest oceanic crust. Geol. Soc. Lond. Spec. Publ. 420 (1), 315–340.
Koptev, A., Burov, E., Gerya, T., Le Pourhiet, L., Leroy, S., Calais, E., Jolivet, L., 2017. Plume-induced continental rifting and break-up in ultra-slow extension context: insights from 3D numerical modeling. Tectonophysics 1–17.
Lee, C.-T.A., 2003. Compositional variation of density and seismic velocities in natural peridotites at STP conditions: Implications for seismic imaging of compositional heterogeneities in the upper mantle. J. Geophys. Res. 108, 2441. https://doi.org/10.1029/2003JB002413.
Levandowski, W., Boyd, O.S., Briggs, R.W., Gold, R.D., 2015. A random-walk algorithm for modeling lithospheric density and the role of body forces in the evolution of the Midcontinent Rift. Geochem. Geophys. Geosyst. 16, 4084–4107.
Li, Z.X., Bogdanova, S.V., Collins, A.S., Davidson, A., De Waele, B., Ernst, R.E., Fitzsimons, I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., 2008. Assembly, configuration, and break-up history of Rodinia: a synthesis. Precambrian Res. 160, 179–210.
Lidiak, E.G., 1996. Geochemistry of subsurface Proterozoic rocks in the eastern Midcontinent of the United States: further evidence for a within-plate tectonic setting. In: van der Pluijm, B.A., Catacosinos, P.A. (Eds.), Basement and Basins of Eastern North America. Geological Society of America Special Paper 308. pp. 45–66.
Lightfoot, P.C., Hawkesworth, C.J., Hergt, J., Naldrett, A.J., Gorbachev, N.S., Fedorenko, V.A., Doherty, W., 1993. Remobilisation of the continental lithosphere by a mantle plume: major-, trace-element, and Sr-, Nd-, and Pb-isotope evidence from picritic and tholeiitic lavas of the Noril'sk District, Siberian Trap, Russia. 114, 171–188. https://doi.org/10.1007/BF00307754.
Liu, M., Stein, S., 2016. Mid-continental earthquakes: spatiotemporal occurrences, causes, and hazards. Earth Sci. Rev. 162, 364–386.
Luza, K.V., Lawson, J.E., 1981. Seismicity and tectonic relationships of the Nemaha Uplift, part III. In: Oklahoma Geological Survey Special Publication, pp. 81–83.
Malone, D.H., Stein, C.A., Craddock, J.P., Kley, J., Stein, S., Malone, J.E., 2016. Maximum depositional age of the Neoproterozoic Jacobsville Sandstone: implications for the evolution of the Midcontinent Rift. Geosphere 12, 1271–1282.
Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J., 2002. Characteristics of volcanic rifted margins. In: Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J. (Eds.), Volcanic Rifted Margins. 362. Geological Society of America Special Paper, Boulder, Colorado, pp. 1–14.
Merdith, A.S., Collins, A.S., Williams, S.E., Pisarevsky, S., Foden, J.D., Archibald, D.B., Blades, M.L., Alessio, B.L., Armistead, S., Plavsa, D., Clark, C., 2017. A full-plate global reconstruction of the Neoproterozoic. Gondwana Res. 50, 84–134.
Merino, M., Keller, G.R., Stein, S., Stein, C., 2013. Variations in Mid-Continent Rift magma volumes consistent with microplate evolution. Geophys. Res. Lett. 40, 1513–1516.
Merle, O., 2011. A simple continental rift classification. Tectonophysics 513 (1–4), 88–95.
Moecher, D.P., Bowersox, J.R., Hickman, J.B., 2018. Zircon U-Pb Geochronology of Two Basement Cores (Kentucky, USA): implications for Late Mesoproterozoic Sedimentation and Tectonics in the Eastern Midcontinent. J. Geol. 126 (1), 25–39.
Moidaki, M., Gao, S., Liu, K.H., Atekwana, E., 2013. Crustal thickness and Moho sharpness beneath the Midcontinent rift from receiver functions. Res. Geophys. 3, 1.
Morgan, W.J., 1971. Convection plumes in the lower mantle. Nature 230 (5288), 42.
Morgan, W.J., 1981. Hotspot tracks and the opening of the Atlantic and Indian Oceans. In: Emiliani, C. (Ed.), The Sea. vol. 7. Wiley, New York, pp. 443–487.
Morgan, W.J., 1983. Hotspot tracks and the early rifting of the Atlantic. Tectonophysics 94, 123–139.
Moulin, M., Aslanian, D., Unternehr, P., 2010. A new starting point for the South and Equatorial Atlantic Ocean. Earth Sci. Rev. 98, 1–37.
Nicholson, S.W., Shirey, S.B., Schulz, K.J., Green, J.C., 1997. Rift-wide correlation of 1.1 Ga Midcontinent rift system basalts, implications for multiple mantle sources during rift development. Can. J. Earth Sci. 34, 504–520.
Ojakangas, R.W., Morey, G.B., 1982. 7A: Keweenawan pre-volcanic quartz sandstones and related rocks of the Lake Superior region. Geol. Soc. Am. Mem. 156, 85–96.
Ojakangas, R.W., Morey, G.B., Green, J.C., 2001. The Mesoproterozoic midcontinent rift system, Lake Superior region, USA. Sediment. Geol. 141, 421–442.
Ola, O., Frederiksen, A.W., Bollmann, T., van der Lee, S., Darbyshire, F., Wolin, E., Revenaugh, J., Stein, C., Stein, S., Wysession, M., 2015. Anisotropic zonation in the lithosphere of Central North America: Influence of a strong cratonic lithosphere on the Mid-Continent Rift. Tectonophysics 683, 367–381.
Richards, M.A., Duncan, R.A., Courtillot, V.E., 1989. Flood basalts and hot-spot tracks: plume heads and tails. Science 246 (4926), 103–107.
Rivers, T., Culshaw, N., Hynes, A., Indares, A., Jamieson, R., Martignole, J., 2012. The Grenville Orogen — a post-LITHOPROBE perspective. In: Percival, J.A., Cook, F.A., Clowes, R.M. (Eds.), Tectonic Styles in Canada: The LITHOPROBE Perspective. Geological Association of Canada Special Paper 49. pp. 97–236.
Roberts, D.G., Bally, A.W. (Eds.), 2012. Regional Geology and Tectonics: Phanerozoic Rift Systems and Sedimentary Basins. Elsevier (500 pp.).
Rooney, T.O., 2017. The Cenozoic magmatism of East-Africa: part I—flood basalts and pulsed magmatism. Lithos 286, 264–301.
Saria, E., Calais, E., Altamimi, Z., Willis, P., Farah, H., 2013. A new velocity field for Africa from combined GPS and DORIS space geodetic solutions, contribution to the definition of the African reference frame (AFREF). J. Geophys. Res. 118, 1677–1697.
Schmandt, B., Lin, F.C., Karlstrom, K.E., 2015. Distinct crustal isostasy trends east and west of the Rocky Mountain Front. Geophys. Res. Lett. 42 (23).
Schnabel, M., Franke, D., Engels, M., Hinz, K., Neben, S., Damm, V., Grassmann, S., Pelliza, H., Dos Santos, P.R., 2008. The structure of the lower crust at the Argentine continental margin, South Atlantic at 44 S. Tectonophysics 454 (1–4), 14–22.
Schutt, D.L., Lesher, C.E., 2006. Effects of melt depletion on the density and seismic velocity of garnet and spinel lherzolite. J. Geophys. Res. 111 (B5).
Sengör, A.M., Burke, K., 1978. Relative timing of rifting and volcanism on Earth and its tectonic implications. Geophys. Res. Lett. 5 (6), 419–421.
Seton, M., et al., 2012. Global continental and ocean basin reconstructions since 200 Ma. Earth Sci. Rev. 113, 212–270. https://doi.org/10.1016/j.earscirev.2012.03.002.
Shay, J., Tréhu, A., 1993. Crustal structure of the central graben of the Midcontinent Rift beneath Lake Superior. Tectonophysics 225 (4), 301–335.
Shen, W., Ritzwoller, M.H., 2016. Crustal and uppermost mantle structure beneath the United States. J. Geophys. Res. Solid Earth 121 (6), 4306–4342.
Shen, W., Ritzwoller, M., Schulte-Pelkum, V., 2013. Crustal and uppermost mantle structure in the central U.S. encompassing the Midcontinent Rift. J. Geophys. Res. 118, 4325–4344.
Simiyu, S.M., Keller, G.R., 1997. An integrated analysis of lithospheric structure across the East African plateau based on gravity anomalies and recent seismic studies. Tectonophysics 278 (1), 291–313.
Stein, S., Wysession, 2003. An Introduction to Seismology, Earthquakes, and Earth Structure. Blackwell.
Stein, S., Van Der Lee, S., Jurdy, D., Stein, C., Wiens, D., Wysession, M., Revenaugh, J., Frederiksen, A., Darbyshire, F., Bollmann, T., Lodewyk, J., 2011. Learning from failure, the SPREE Mid-Continent Rift Experiment. GSA Today 21, 5–7.
Stein, C.A., Stein, S., Merino, M., Keller, G.R., Flesch, L.M., Jurdy, D.M., 2014. Was the Mid-continent Rift part of a successful seafloor-spreading episode? Geophys. Res. Lett. 41, 465–1470.
Stein, C.A., Kley, J., Stein, S., Hindle, D., Keller, G.R., 2015. North America's Midcontinent Rift: when rift met LIP. Geosphere 11, 1607–1616.
Stein, S., Stein, C., Kley, J., Keller, R., Merino, M., Wolin, E., Jurdy, D., Wiens, D., Wysession, M., Al-Equabi, G., Shen, W., Frederiksen, A., Darbyshire, F., Waite, G., Rose, W., Vye, E., Rooney, T., Moucha, R., Brown, E., 2016. When rift met LIP: new insights about the Midcontinent Rift. Eos 97, 10–16.
Stein, C.A., Stein, S., Elling, R., Keller, G.R., Kley, J., 2018. Is the "Grenville Front" in the central United States really the Midcontinent Rift? GSA Today 28. https://doi.org/10.1130/GSATG357A.
Swanson-Hysell, N.L., Burgess, S.D., Maloof, A.C., Bowring, S.A., 2014. Magmatic activity and plate motion during the latent stage of Midcontinent Rift development. Geology 42. https://doi.org/10.1130/G35271.1.
Thomas, M.D., Teskey, D.J., 1994. An interpretation of gravity anomalies over the Midcontinent Rift, Lake Superior, constrained by GLIMPCE seismic and aeromagnetic data. Can. J. Earth Sci. 31, 682–697. https://doi.org/10.1139/e94-061.
Thomas, W.A., Tucker, R.D., Astini, R.A., Denison, R.E., 2012. Ages of pre-rift basement and synrift rocks along the conjugate rift and transform margins of the Argentine Precordillera and Laurentia. Geosphere 8, 1366–1383.
Thybo, H., Artemieva, I.M., 2013. Moho and magmatic underplating in continental lithosphere. Tectonophysics 609, 605–619.
Thybo, H., Nielsen, C.A., 2009. Magma-compensated crustal thinning in continental rift zones. Nature 457 (7231), 873–876.
Tohver, E., Teixeira, W., van der Pluijm, B., Geraldes, M.C., Bettencourt, J.S., Rizzotto, G., 2006. Restored transect across the exhumed Grenville orogen of Laurentia and Amazonia, with implications for crustal architecture. Geology 34, 669–672.
Trestrail, K.R., Rooney, T.O., Girard, G., Svoboda, C., Yirgu, G., Ayalew, D., Keppelman, J., 2017. Sub-continental lithospheric mantle deformation in the Yerer-Tullu Wellel Volcanotectonic Lineament: a study of peridotite xenoliths. Chem. Geol. 455, 249–263.
Van Schmus, W.R., Hinze, W.J., 1985. The midcontinent rift system. Annu. Rev. Earth Planet. Sci. 13 (1), 345–383.
Vervoort, J.D., Wirth, K., Kennedy, B., Sandland, T., Harpp, K.S., 2007. The magmatic evolution of the Midcontinent rift: New geochronologic and geochemical evidence from felsic magmatism. Precambrian Res. 157 (1–4), 235–268.
Walker, D., Misra, K.C., 1992. Tectonic significance of basalts of the Middle Run Formation (Upper Proterozoic), of the East Continent Rift Basin, Indiana and Kentucky. In: Geological Society of America Abstracts With Programs. 24. pp. 330.
Watts, A.B., Cox, K.G., 1989. The Deccan Traps, an interpretation in terms of progressive lithospheric flexure in response to a migrating load. Earth Planet. Sci. Lett. 93, 85–97. https://doi.org/10.1016/0012-821X(89)90186-6.
White, R.S., 1997. Mantle temperature and lithospheric thinning beneath the Midcontinent rift system, evidence from magmatism and subsidence. Can. J. Earth Sci. 34, 464–475. https://doi.org/10.1139/e17-038.
White, R.S., Mckenzie, D.P., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. J. Geophys. Res. 94, 7685–7730.
Whitmeyer, S.J., Karlstrom, K.E., 2007. Tectonic model for the Proterozoic growth of North America. Geosphere 3 (4), 220–259.

van Wijk, J.W., Huismans, R.S., Ter Voorde, M., Cloetingh, 2001. Melt generation at volcanic continental margins: no need for a mantle plume? Geophys. Res. Lett. 28, 3995–3998.
van Wijk, J.W., van der Meer, R., Cloetingh, S.A.P.L., 2004. Crustal thickening in an extensional regime: application to the mid-Norwegian Vøring margin. Tectonophysics 387, 217–228.
Wolin, E., van der Lee, S., Bollmann, T.A., Wiens, D.A., Revenaugh, J., Darbyshire, F.A., Frederiksen, A.W., Stein, S., Wysession, M.E., 2015. Seasonal and diurnal variations in long-period noise at SPREE stations: the influence of soil characteristics on shallow stations' performance. Bull. Seismol. Soc. Am. 105 (5), 2433–2452.
Wunderman, R.L., Wannamaker, P.E., Young, C.T., 2018. Architecture of the hidden Penokean terrane suture and Midcontinent rift system overprint in eastern Minnesota and western Wisconsin from magnetotelluric profiling. Lithosphere 10 (2), 291–300.
Zhang, H., Lee, S., Wolin, E., Bollmann, T.A., Revenaugh, J., Wiens, D.A., Frederiksen, A.W., Darbyshire, F.A., Aleqabi, G.I., Wysession, M.E., Stein, S., 2016. Distinct crustal structure of the North American Midcontinent Rift from P wave receiver functions. J. Geophys. Res. Solid Earth 121. https://doi.org/10.1002/2016JB013244.
Ziegler, P., Cloetingh, S., 2004. Dynamic processes controlling evolution of rift basins. Earth-Sci. Rev. 64, 1–50.
Zietz, I., King, E.R., Geddes, W., Lidiak, E.G., 1966. Crustal study of a continent strip from the Atlantic Ocean to the Rocky Mountains. Geol. Soc. Am. Bull. 77, 1427–1448.