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North America's Midcontinent Rift: When rift met LIP

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ABSTRACT

Rifts are segmented linear depressions that are filled with sedimentary and igneous rocks; they form by extension and often evolve into plate boundaries. Flood basalts, a class of large igneous provinces (LIPs), are broad regions of extensive volcanism formed by sublithospheric processes. Typical rifts are not filled with flood basalts, and typical flood basalts are not associated with significant crustal extension and faulting. North America's Midcontinent Rift (MCR) is an unusual combination, because its 3000-km length formed during a continental breakup event 1.1 Ga, but it contains an enormous volume of igneous rocks that are mostly flood basalt. We show that MCR volcanic rocks are significantly thicker than other flood basalts, due to their deposition in a narrow rift rather than across a broad region, giving the MCR a rift's geometry but a LIP's magma volume. Structural modeling of seismic-reflection data shows that LIP volcanics were deposited during two phases—an initial rift phase where flood basalts filled a fault-controlled extending basin and a postrift phase where LIP volcanics and sediments were deposited in a thermally subsiding sag basin without associated faulting. The crust thinned during the initial rifting phase and then rethickened during the postrift phase and later compression, yielding the present thicker crust observed seismologically. The restriction of extension to a single normal fault in each rift segment, steeply inward-dipping rift shoulders with sharp hinges, and persistence of volcanism after rifting ended gave rise to a deep flood basalt–filled rift geometry not observed in other presently active or ancient rifts. The unusual coincidence of a rift and LIP arose when a new rift associated with continental breakup interacted with a mantle plume or overrode anomalously hot or fertile upper mantle.

■ INTRODUCTION

One of the most prominent features on gravity and magnetic maps of North America is the Midcontinent Rift (MCR), an extensive band of buried igneous and sedimentary rocks that outcrop around Lake Superior (Fig. 1A). To the

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south, the rift is deeply buried by younger sediments, but its two arms are easily traced because the igneous rocks are dense and highly magnetized (King and Zietz, 1971; Hinze et al., 1997). The MCR formed at ca. 1.1 Ga within Laurentia, the Precambrian core of the North American continent, by extension and volcanism followed by subsidence and sedimentation. It is considered a rift because of its morphology as a long fault-bounded and segmented depression filled by volcanic rocks and sediments. It appears to have formed as part of the rifting of Amazonia (Precambrian northeast South America) from Laurentia (Precambrian North America) and became inactive once seafloor spreading was established (Stein et al., 2014). Hence, it can be viewed as an analogue to today's East African Rift system along which the Nubian and Somalian plates are diverging (Saria et al., 2013), which contains microplates with boundaries analogous to the MCR's two arms (Merino et al., 2013). It can also be viewed as a preserved piece of what might have evolved to a volcanic passive continental margin (Roberts and Bally, 2012).

The MCR is an unusual combination of a rift and a flood basalt (Green, 1983), two major types of features associated with continental volcanism that differ in geometry and origin (Foulger, 2011). Rifts are segmented linear depressions filled with sedimentary and igneous rocks, which form by extension and often evolve into plate boundaries (Roberts and Bally, 2012). Flood basalts, a class of large igneous provinces (LIPs), are broad regions of extensive volcanism formed by sublithospheric processes (Ernst, 2014). Typical rifts are not associated with flood basalts, and typical flood basalts are not associated with significant crustal extension and faulting. However, the MCR is a 3000-km-long rift formed as part of a continental breakup event 1.1 Ga; but it contains an enormous volume of igneous rocks, mostly flood basalt typical of a LIP.

The prominent positive Bouguer anomaly characterizing the MCR due to the high-density volcanics filling it (Fig. 1B) illustrates its unusual nature. In contrast, typical continental rifts, such as the Rio Grande rift in the western United States, have negative gravity anomalies (Fig. 1C) because they are largely filled with low-density sediment, whose effects overwhelm that of the higher density mantle at shallow depth due to crustal thinning that occurred during the extension. Modeling of seismic and gravity profiles across the MCR indicates a total magma volume of ~1–2 × 10^6 km^3 (Hutchinson et al., 1990; Merino et al., 2013), well above the large LIP threshold of 10^5 km^3 (Ernst, 2014).
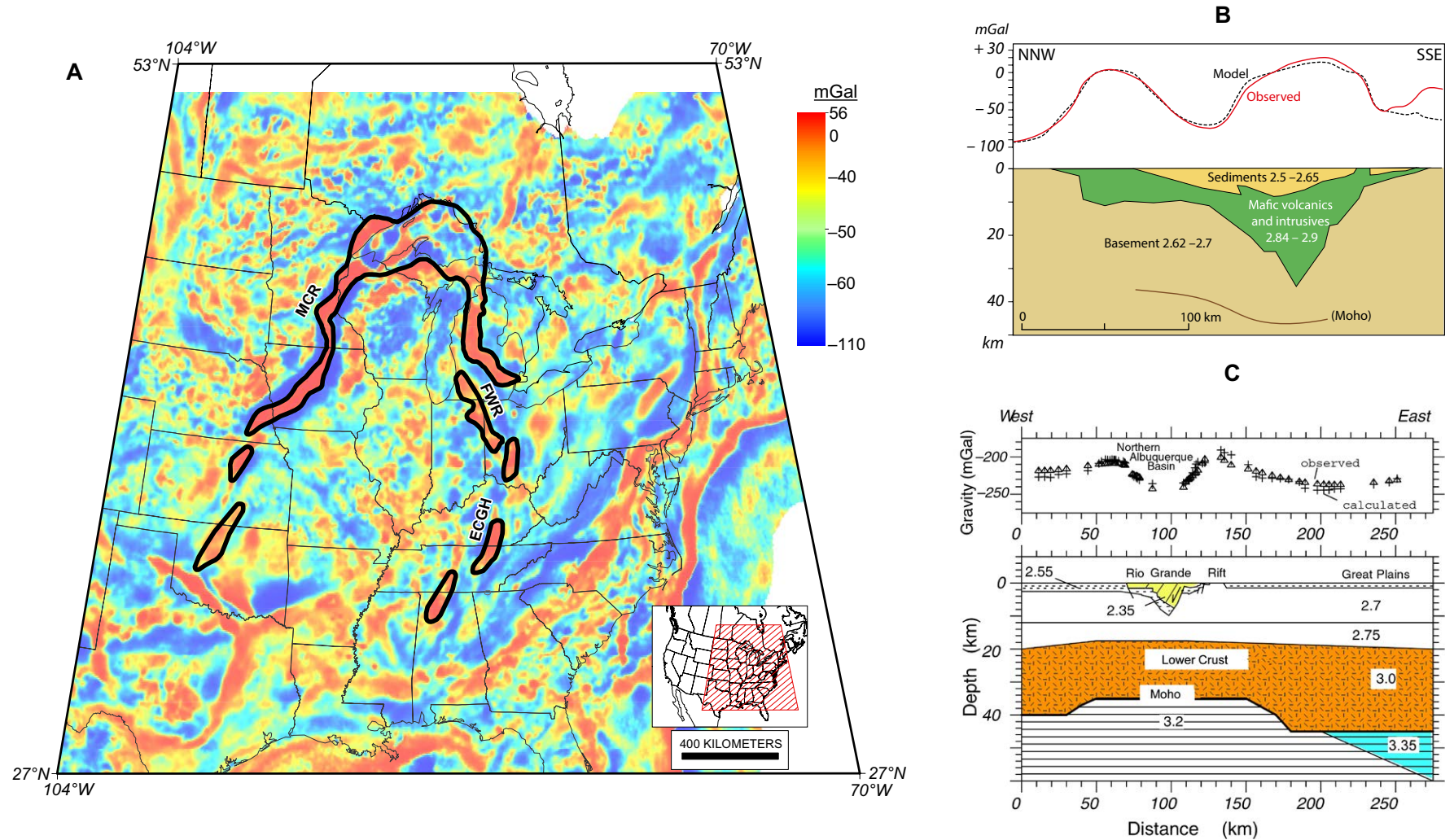
Figure 1. (A) Gravity map showing Midcontinent Rift (MCR), and its extensions, the Fort Wayne Rift (FWR) and East Continent Gravity High (ECGH), computed by upward continuing complete Bouguer anomaly (CBA) data to 40 km and subtracting result from CBA (Stein et al., 2014). (B) Bouguer gravity data and model across the MCR, showing positive anomaly due to high-density volcanics. The black dashes are calculated (model) gravity, and the red line is observed gravity (Thomas and Teskey, 1994). (C) Bouguer gravity data and integrated geophysical model across the Rio Grande rift showing negative anomaly due to low-density sediments (Grauch et al., 1999).

■ COMPARISON WITH OTHER LIPS

The MCR has more magma than many classic flood basalts, including the Columbia River basalts and Deccan Traps (Fig. 2). Comparing its volume and surface area to those of other flood basalts 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 magma volume and composition of a LIP.

■ STRUCTURAL ANALYSIS

The architecture of the MCR is shown by the Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) seismic-reflection lines across western Lake Superior (Green et al., 1989). The profiles, such as line C (Fig. 3A), 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 volcanic rocks 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. Thinner volcanic and overlying sedimentary successions occur beyond the faults on both flanks. The upper regions of the faults show reverse-sense offsets of stratigraphic markers, due to basin inversion (Chandler et al., 1989) long after rifting, volcanism, and subsidence ended.

The reflection data indicate a combined history of extension, volcanism, subsidence, and reverse faulting (Cannon, 1992) whose sequence and magnitude we constrain by stepwise structural reconstruction. No depth-converted versions of the GLIMPCE seismic lines have been published, but some approximate depth markers are shown on interpreted time sections (Green et al.,

1989). We used these depth markers to adjust the vertical scale of the structural models. The associated uncertainties do not affect our interpretation of the evolution and have little effect on our estimates of extension and shortening magnitudes.

Examination of line C shows that 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.

We model this history via a numerical stepwise structural restoration, working backward from the present geometry. The stepwise restorations of cross sections (Figs. 3 and 4) were carried out using Midland Valley´s 2DMove software. The reverse offsets on the bounding faults were removed by joining the footwall and hanging-wall cutoffs of the base of the postrift sedimentary succession.

TABLE 1. DATA AND SOURCES FOR FIGURE 2

Continental flood basalt	Area (10⁶ km²)	Volume (10⁶ km³)
Columbia River Basalts[1]	0.21	0.21
Ethiopian Traps[2]	0.75	0.45
Deccan Traps[3]	1.1	1.3
North Atlantic Igneous Province (NAIP)[4]	1.3	1.8
Paraná-Etendeka[5]	2.0	2.2
Karoo-Ferrar[6]	2.7	2.7
Central Atlantic Magmatic Province (CAMP)[7]	10.0	3.15
Siberian Trap[8]	7.0	4.0
Emeishan[9]	0.25	0.35
Midcontinent Rift (MCR)[10]	0.36	2.1

[1]Area and volume from Reidel et al. (2013).
[2]Only for Africa; area and volume from Mohr and Zanettin (1988).
[3]Area and volume from Jay and Widdowson (2008).
[4]Area and volume from Eldholm and Grue (1994).
[5]Area from Ernst (2014) and volume from Marks et al. (2014).
[6]Area from Klausen (2009) and Ross et al. (2005) and volume from Storey and Kyle (1997) and Storey et al. (2013).
[7]The igneous estimates for CAMP include tholeiitic dikes and sills, with smaller amounts of flood basalts (Bensalah et al., 2011); so this may explain the outlier position. Area from Marzoli et al. (2014) and volume from Bensalah et al. (2011).
[8]Includes the areas of the Siberian platform and the west Siberian Basin; so includes more than just flood basalts; volume estimates range from 2.3 to 4 × 10⁶ km³; so average value used. Area and volume from Ernst (2014).
[9]Volume estimates range from 0.3 to 0.4 10⁶ km³; so average value used. Area and volume from Ali et al. (2010).
[10]MCR area and volume using our recalculated values for parts of the west (Woelk and Hinze, 1991) and east arms (Zhu and Brown, 1986) and new lengths from Stein et al. (2014) and from Chandler et al. (1989), Hinze et al., (1990), and Hutchinson et al. (1990).
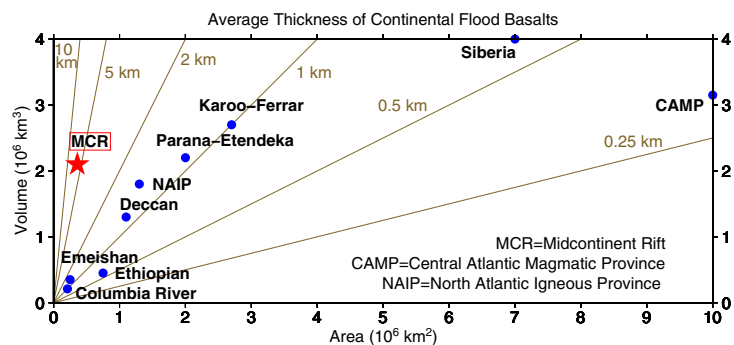

Figure 2. Comparison of volume, area, and average thickness for various continental flood basalts. The Midcontinent Rift (MCR) volcanics are comparable in volume to other flood basalts but thicker because they are deposited in a narrow rift. Data are tabulated in Table 1.
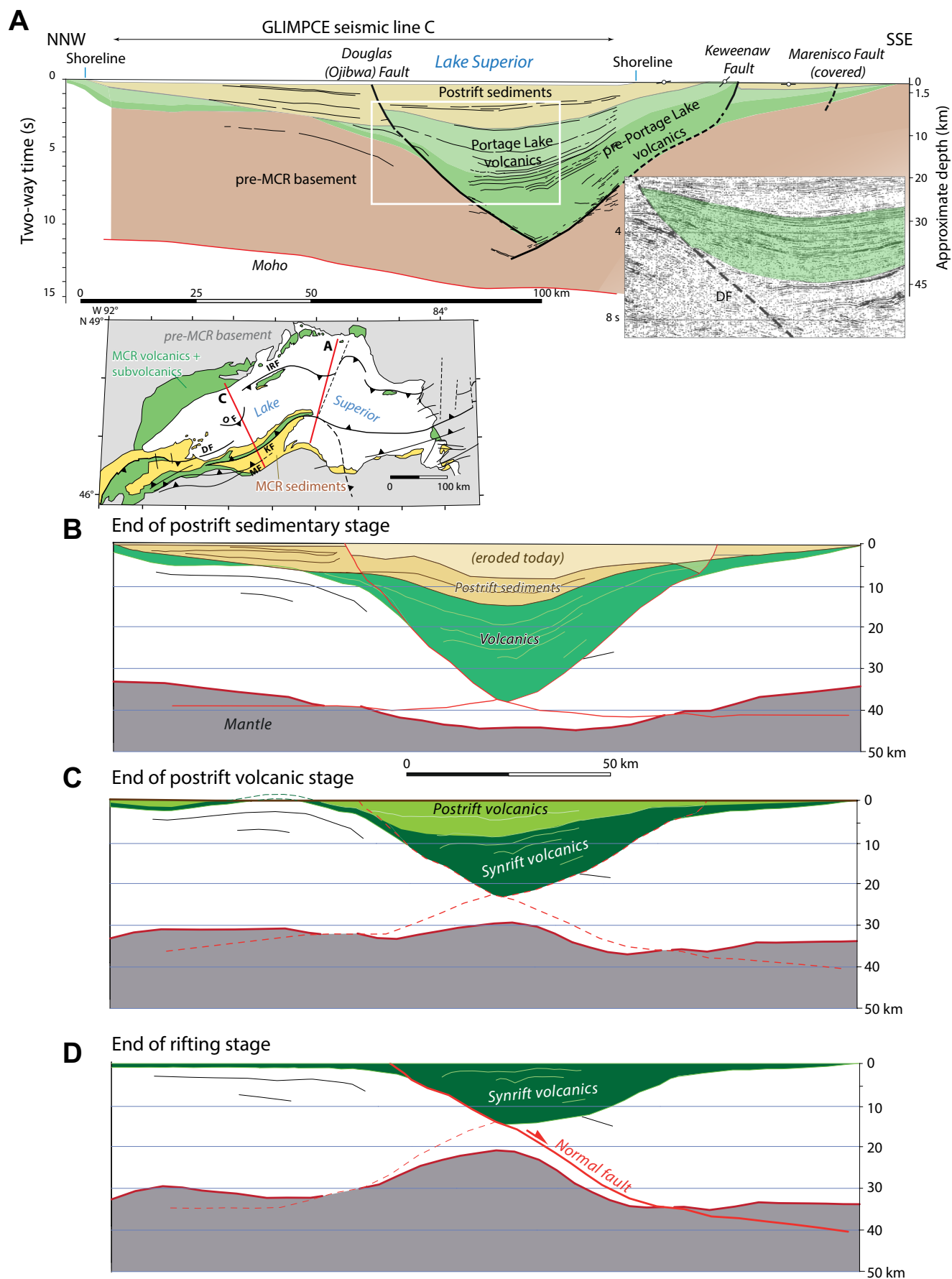
Figure 3. (A) Geologic cross section of the Midcontinent Rift (MCR) based on a line drawing of Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) seismic line C (slightly modified from Green et al., 1989) and complemented with land data in the south, showing geometry of Portage Lake volcanic rocks and postrift Oronto sediments in the rift basin. Left inset is tectonic sketch map (Manson and Halls, 1997) with major faults and locations of GLIMPCE lines C and A. Because the Ojibwa fault was treated as part of the Douglas fault until recently, we use both names. Abbreviations are DF—Douglas fault; OF—Ojibwa fault; IRF—Isle Royale fault; KF—Keweenaw fault; MF—Marenisco fault. Right inset is close-up of migrated line C (Milkereit et al., 1990). Postrift volcanics (green) dip and thicken southward. (B) to (D) Stepwise restoration of section shown in (A). (B) Reverse offsets on bounding faults removed; top of postrift sediments made horizontal. (C) Top of postrift volcanics made horizontal. (D) Top of synrift volcanics made horizontal. Synrift volcanics fill a half graben bounded to the north by Douglas-Ojibwa listric normal fault.

GLIMPCE Line A

A After Inversion (Present)

N   Isle Royale Fault   Keweenaw Fault   S

postrift sediments

Volcanics

Dense lower crust

Mantle

0
10
20
30
40
50
60 km

B End of postrift stage

0   50   100 km

(eroded today)

postrift volcanics

synrift volcanics

Future Isle Royale Fault

0
10
20
30
40
50
60 km

C End of rifting stage

synrift volcanics

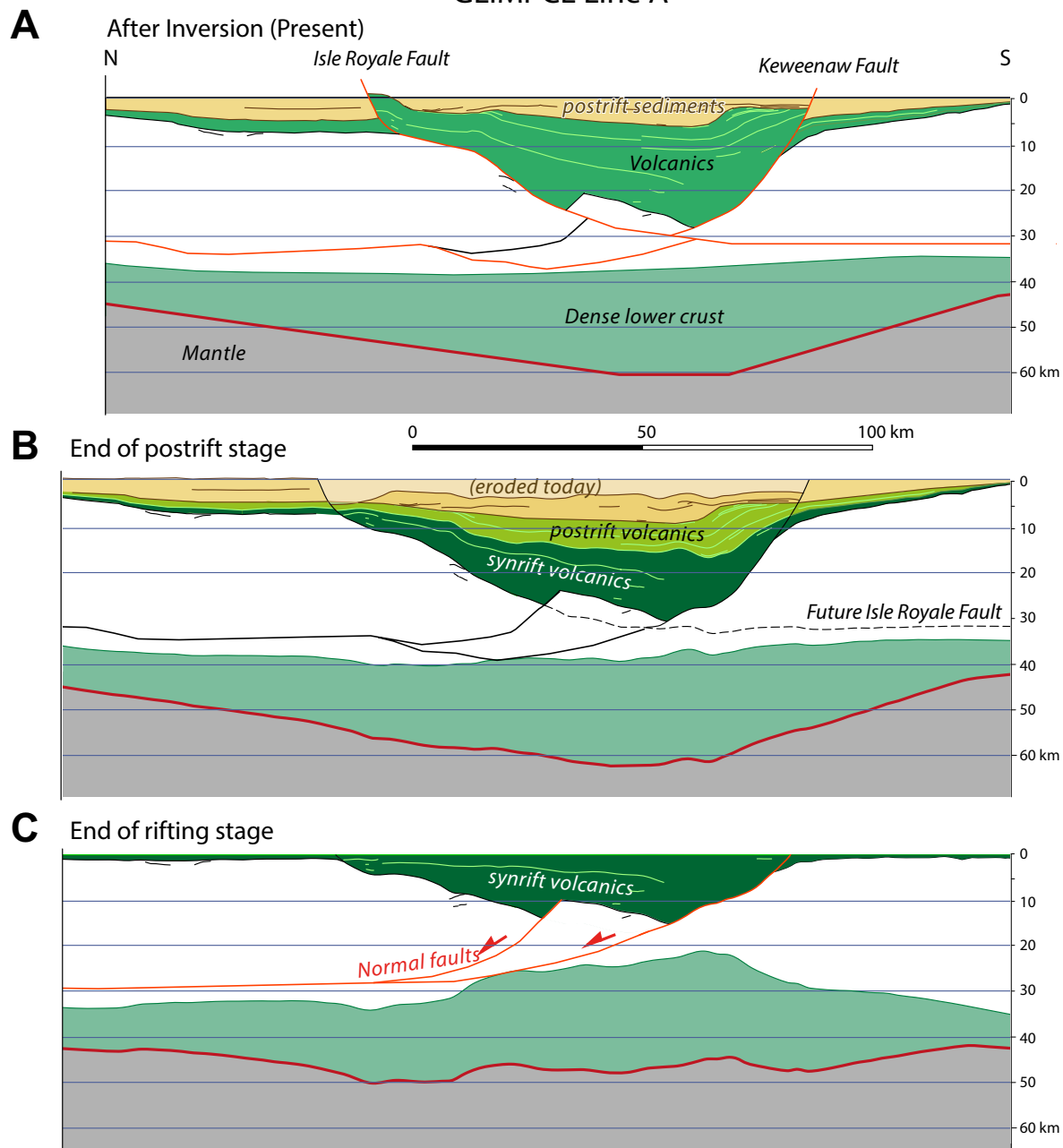Normal faults

0
10
20
30
40
50
60 km

Figure 4. Stepwise restoration of geologic cross section based on line drawing of Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) line A (slightly modified from Green et al., 1989). (A) Present situation after inversion. (B) Reverse offsets on bounding faults removed; top postrift sediments horizontal. (C) Top synrift volcanics horizontal. Synrift volcanics fill a half graben bound to the Keweenaw listric normal fault and an associated splay fault. Notice mirror-image configuration compared to GLIMPCE line C (Fig. 3).

For the Keweenaw fault in section C, the faults and the marker horizon had to be extrapolated a short distance above the present-day surface to reconstruct the location of the eroded hanging-wall cutoffs. Elements in the hanging walls were restored to their pre-inversion locations by the fault-parallel flow algorithm that moves all points on trajectories parallel to the faults.

All other restoration steps until the situation at the end of rifting were performed using vertical simple shear. The procedure involves the projection of a marker line onto a gently curved (Fig. 3B) or horizontal target line (all other steps) along vertical paths. All points located on the same vertical line are moved the same distance and in the same direction. Cross-section area is conserved.

The first step removes the postrift reverse offsets on the bounding faults, making the base of the postvolcanic sedimentary succession continuous (Fig. 3B). This yields a lenticular sedimentary body with a flat top and sagging base, with thickness greatest above the MCR axis and tapering to both sides. The second step (Fig. 3C) removes the postrift sediments and restores the top of the volcanic sequence to a horizontal surface, consistent with the depositional geometry of low-viscosity lava flows.

The upper part of the volcanic succession now has a lenticular geometry similar to that of the postrift sediments with no evidence of syndepositional faulting, indicating deposition during the postrift stage. Particularly on the southern flank, the dip of the basement steepens abruptly across a "hinge." The lower part of the volcanic succession is asymmetric, thickening with increasing dip toward the north, where reflections truncate against the south-dipping continental basement, suggesting a fault contact. The third step (Fig. 3D) restores the base of the postrift volcanics to horizontal, representing the situation at the end of rifting. The synrift volcanics show a wedge shape typical of a half-graben fill overlying a listric normal fault. 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 ~10 m.y. (half the volcanic succession) gives an extension rate of ~2–2.5 mm/yr, a typical value for rifts.

Line C shows that the Douglas-Obijwa 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 (Hinze et al., 1997), indicating it was not a large rift-bounding normal fault during the extensional phase. In contrast, GLIMPCE line A to the east (Fig. 4) shows that the Keweenaw fault was the master normal fault with 28 km of extension. This polarity reversal along a series of adjacent half-graben segments (Fig. 5) (Sexton and Henson, 1994; Dickas and Mudrey, 1997) is analagous to that observed in the East African rift.

We estimated the amount of extension using forward models of the half-graben infill (Fig. 5). These were created using the "Horizons from fault" function in 2DMove, with hanging- wall deformation again set to vertical simple shear. We employed trial-and-error fitting of fault geometries and extension magnitudes until reasonable agreement was obtained between modeled and observed geometries, using the base of the synrift volcanics as primary marker. The addition of "syntectonic beds" also allows us to compare the model prediction to observed reflection patterns within the synrift volcanics.

■ TECTONIC IMPLICATIONS

The restorations show that the MCR began as a half graben that was filled by synrift flood basalts during the rift phase. After extension ceased, it further subsided and accommodated another thick succession of flood basalts during the postrift phase. After LIP volcanism ended, thermal subsidence continued, accompanied by postrift sedimentation. The crust was depressed and strongly flexed (Peterman and Sims, 1988) deepening the Moho under the load of the dense flood-basalt infill and sediment. This geometry is remarkably different from that observed at other LIPs, where the stacked basalt flows without significant overlying sediment produced only minor flexure (Watts and Cox, 1989). The MCR crust beneath the rift and particularly at the "hinges" may have progressively weakened during deposition as it flexed (Ranalli, 1994), causing greater subsidence and contributing to the unusually deep basin.

Although the volcanic rocks are well dated, the onset of extension is not. We suspect that it began ca. 1120 Ma, coincident with the end of strike-slip motion of Amazonia as it separated from Laurentia (Tohver et al., 2006; Stein et al., 2014). This is after regional volcanism began at ca. 1150 Ma but before the ca. 1109 Ma MCR flood basalts (Heaman et al., 2007). The bulk of the pre–Portage Lake volcanics—equivalent to the "Early magmatic stage" (Miller and Nicholson, 2013)—was accommodated in the active half graben of the rift phase. The switch to the overlying Portage Lake flows is marked by continuous reflections on the seismic lines and coincides with the transition to broad subsidence above and beyond the aborted graben. The end of extension and the following postrift subsidence phase thus coincide with the "latent" and "main magmatic stages" of the MCR. The volcanics of both the early and main magmatic stages have been interpreted as largely mantle plume melts (Nicholson et al., 1997; White, 1997).

Extension ended about the time the Portage Lake deposition began (ca. 1096 Ma) (Davis and Paces, 1990; Nicholson et al., 1997); so a significant portion of the volcanics was not deposited during extension. Similarly, extension ended long before the regional compression that inverted the basin by reverse 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.

The reconstructions show how crustal thickness—defined as the depth of the Moho—evolved. The original crust was thinned in the rifting stage. It then rethickened during the postrift phase and was thickened further (by ~5 km for line C) by reverse faulting during the later basin inversion. Thickening by reverse faulting for line A was ~6 km. Additional thickening may also have occurred via magmas added to the base of the crust. The crustal thickening in GLIMPCE lines A and C is consistent with observations of crustal thickening elsewhere along the MCR (Shen et al., 2013).

Why line A has ~40% more extension and 40% more compression than line C is not certain. The directions of extension and of the subsequent multiple episodes of compression are difficult to determine independently. Because
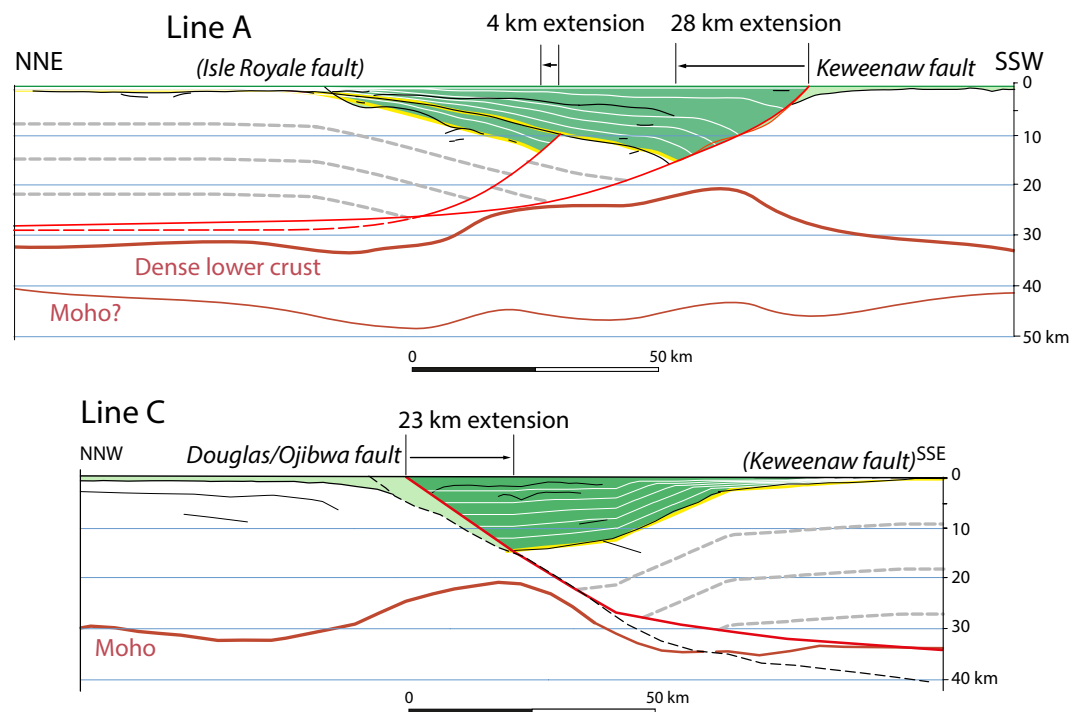
Figure 5. Area-balanced forward models of the half-graben fills in Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) lines A and C (green, with thick yellow baselines and white layering) compared to line drawing (thin black lines). Pale-green areas are volcanics outside the modeled half-graben fills. Fault segments in parentheses are inactive during extension.

the trends of the two lines are ~35° apart, ~22% may be a geometrical effect assuming the extension and shortening directions were the same for the two lines. Difference in pre-rift structures and thus mechanical strength could have influenced the extension amounts, and the greater extension for line A could have resulted in a weaker region and more compression.

Our scenario for the sequence of rift extension and then LIP deposition for the Lake Superior portion of the MCR, summarized schematically in Fig. 6, is based on high-quality marine seismic-reflection data plus direct sampling and geochronological dating of the exposed volcanic rocks. We suspect the remainder of the MCR behaved similarly but cannot confirm this for the buried west and east arms of the MCR because comparable data are not available. However, the available seismic data (Zhu and Brown, 1986; Chandler et al., 1989; Woelk and Hinze, 1991) and older petrological analysis from a few drill-hole samples (Keller et al., 1982; Dickas et al., 1992; Walker and Misra, 1992; Cullers and Berendsen, 1993; Lidiak, 1996) show similarities to what is observed in the Lake Superior area.

The MCR is unusual in hosting the world's largest deposit of native copper (copper not bounded to other elements) as well as copper sulfide deposits like those found elsewhere. Native copper deposits are not associated with other rifts or LIPs. Bornhorst (1997) and Bornhorst and Barron (2011) attribute them to volatile degassing creating sulfur-deficient flood basalts, followed by hydrothermal fluids leaching copper from the thick basalts at temperatures of 300–500 °C after rifting ceased, and rising in the permeable pathways primarily provided by the reverse faults. This situation arose because the MCR's combination of a rift and LIP gave rise to unusually thick basalts buried under thick sediments that kept the basalt at high temperatures, allowing extraction of large amounts of copper.

The reconstruction shows how the MCR's unusual architecture of a very deep rift with a large-volume flood basalt evolved from the coincidence of a rift with a LIP. This combination resolves the paradox that the rifting requires tectonic stresses and faulting atypical of LIPs but consistent with coeval continental breakup (Stein et al., 2014), whereas the volume and composition of the volcanic rocks are interpreted as showing that the MCR formed over a deep-seated mantle plume (Nicholson et al., 1997; White, 1997). The reconstruction demonstrates that both occurred. A mantle plume impinging upon lithosphere under extension has been hypothesized (Courtillot et al., 1999) and simulated numerically (Burov and Gerya, 2014). Because our analysis is based on structures within the crust, it does not require specific aspects of those scenarios.

North America's Midcontinent Rift Evolution

Rifting (extension) begins

Crust (Basement rocks)

Mantle

About 1120–1109 Ma

Rifting and volcanism, normal faults active, crustal thinning, Pre–Portage Lake volcanics

About 1109–1096 Ma

Subsidence and volcanism, faults inactive, crustal thickening, Portage Lake volcanics

About 1096–1086 Ma

Subsidence and sedimentation, faults inactive, crustal thickening

About 1086–? Ma

SSE

Compression, reverse faulting and uplift, additional crustal thickening

Much later

NNW   Douglas (Ojibwa) Fault   Lake Superior   Shoreline   Keweenaw Fault   SSE

0
10
20   Present
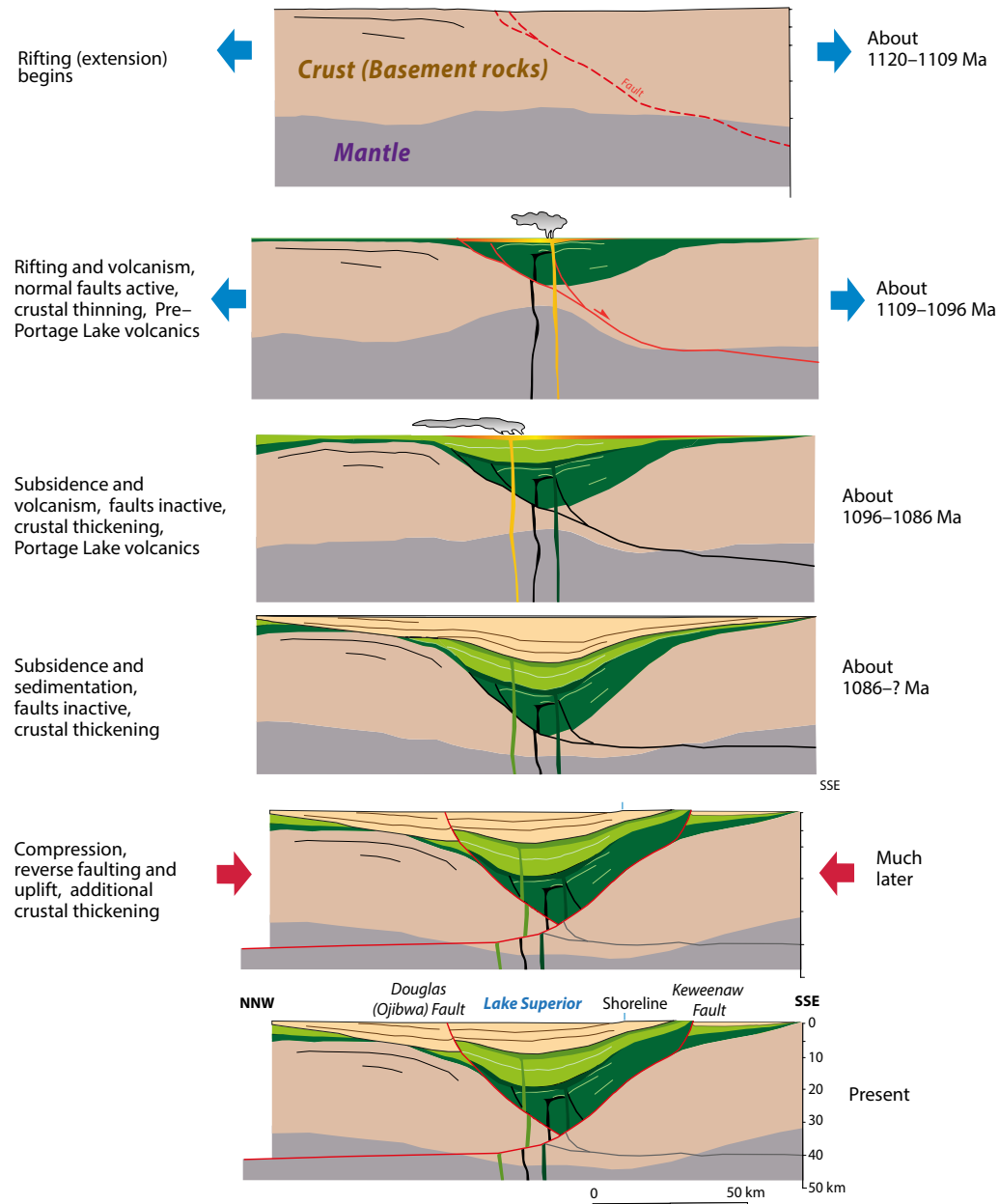30
40
50 km

0   50 km

Figure 6. Schematic evolution of the Midcontinent Rift.

However, the combination of rifting and LIP volcanism in the MCR implies a scenario in which a rifting continent by chance overrode a plume or a shallow region of anomalously hot or fertile upper mantle (Silver et al., 2006). It is worth noting that it is still unclear how the magma source operated over a long period of rapid plate motion (Swanson-Hysell et al., 2014).

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