LETTERS

The importance of rift history for volcanic margin formation

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Rifting and magmatism are fundamental geological processes that shape the surface of our planet. A relationship between the two is widely acknowledged but its precise nature has eluded geoscientists and remained controversial. Largely on the basis of detailed observations from the North Atlantic Ocean, mantle temperature was identified as the primary factor controlling magmatic production[1], with most authors seeking to explain observed variations in volcanic activity at rifted margins in terms of the mantle temperature at the time of break-up[2,3]. However, as more detailed observations have been made at other rifted margins worldwide, the validity of this interpretation and the importance of other factors in controlling break-up style have been much debated[4–7]. One such observation is from the northwest Indian Ocean, where, despite an unequivocal link between an onshore flood basalt province, continental break-up and a hot-spot track leading to an active ocean island volcano, the associated continental margins show little magmatism[5,8]. Here we reconcile these observations by applying a numerical model that accounts explicitly for the effects of earlier episodes of extension. Our approach allows us to directly compare break-up magmatism generated at different locations and so isolate the key controlling factors. We show that the volume of rift-related magmatism generated, both in the northwest Indian Ocean and at the better-known North Atlantic margins, depends not only on the mantle temperature but, to a similar degree, on the rift history. The inherited extensional history can either suppress or enhance melt generation, which can explain previously enigmatic observations.

Extensive surveys in the North Atlantic have shown that the breakup of Europe from North America was characterized by massive episodes of igneous activity. Such large magmatic events are potentially highly significant to the Earth's history, and have been linked to both mass extinctions and changes in global climate[9]. Magmatically enhanced (volcanic) margins extend laterally over distances of 2,000 km with spatial and temporal relationships suggesting a direct link to the Iceland hot spot[10]. These first-order observations were explained[1] using a model of melt generation due to uniform, instantaneous extension. The calculations[1] showed little dependence on the initial lithospheric thickness when full break-up occurred and it was concluded that the primary constraint on magmatism was mantle temperature. A later study showed that only for very slow extensional rates would the relationship between high mantle temperature and voluminous magmatism break down[11]. It has therefore been widely assumed that all hot spots associated with continental break-up will generate a volcanic margin, with the volume of melt decreasing steadily along strike away from the hot-spot centre[12]. Subsequent challenges to this idea have centred on examples of volcanic margins where the presence of a hot spot at the time of break-up is disputed[4,7] or where rapid along-strike variations in magmatism are observed[6,13,14]. Here we focus on new observations from the northwest Indian Ocean, where, like in the North Atlantic, there is an unequivocal association between continental break-up, an onshore flood basalt province (the Deccan Traps) and a hot-spot track (the Chagos–Laccadive–Mascarene ridge) leading to an active volcanic island (Réunion). As the size of the Deccan Traps is similar to that of the North Atlantic igneous province, it has been widely inferred that the associated continental margins should also display volcanic characteristics[12,15].

Continental break-up is commonly preceded by a series of extensional episodes[16], which have an effect on the subsequent volcanic nature of the rifted margin. However, such geological considerations are not generally incorporated into numerical models of melt generation during extension. Instead previous researchers implicitly assumed a geometry of the lithosphere[1,11], imposed the geometry of the rift location as an initial condition[2,3] or assumed the lithosphere to have uniform vertical thickness with a predefined weakness[17]. We previously developed a self-consistent numerical modelling approach that incorporates multiple episodes of extension suggested by geological data[18]. Our early model outputs were tested against observations from southwest Greenland, where the extension rate was low (half-spreading rate, $<40\ \mathrm{mm\,yr^{-1}}$), and suggested that at this particular margin extension before break-up thinned the lithosphere and that this thin region aided the decompression of hot upwelling mantle. Here we build upon this work by applying our model to the conjugate margin in the eastern Atlantic and to rifting between India and the Seychelles, where extension was fast (half-spreading rate, $>60\ \mathrm{mm\,yr^{-1}}$). We show that mantle thermal structure alone is unable to explain the observations in either the northwest Indian Ocean or in the North Atlantic. Rather, it is the geological inheritance from previous episodes of extension that controls the volcanic nature of break-up. Therefore, our work provides a new framework in which to understand the structure of continental margins worldwide.

The key advance of our model is that as it evolves through time, it tracks melt generation and the associated depletion of the solid mantle, allowing for a better representation of the inherited melting conditions (see Methods Summary and Supplementary Information for more details). We solve the equations of thermal convection for a non-Newtonian viscous fluid using an extended version of Citcom[3,18,19], which tracks the melt and solid-mantle major-element composition throughout the lithosphere as melting progresses[18]. Melt generation results from decompression as the solid mantle crosses the solidus, which is a function of depth and depletion. The volume and seismic velocity of any melt generated through time is then calculated and compared with measurements from wide-angle seismic experiments. To assess the influence of hot mantle, a thermal anomaly beneath the lithosphere is included as a layer of predefined thickness and excess temperature. In the analysis presented here, this layer is not replenished, because we model distal volcanic margins where observations suggest that the excess magmatism takes the form of an initial pulse that decays

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once sea-floor spreading is under way[3]. Extension that leads to break-up is imposed in our dynamic model by applying divergent surface boundary conditions. We incorporate the multiple episodes of extension by shifting the centre of the divergent boundary condition on the basis of geological constraints, allowing the lithosphere to cool by conduction between periods of extension.

In 2003, we collected conjugate wide-angle seismic profiles across the Seychelles margin and the Indian margin in a region known as the Laxmi ridge, which were approximately 1,000 km from the Deccan Traps at the time of break-up (Fig. 1a). However, rather than a heavily intruded continent–ocean transition zone and thick oceanic crust, only a modest addition of magma within the transition zone and relatively thin oceanic crust was found[5,8]. Rare-earth-element inversions suggest that mantle temperatures during the eruption of the Deccan Traps were up to 200 °C higher than the background temperatures[15]. If the Seychelles/Laxmi ridge margin inherited such a thermal structure, our model predicts that rifting of a previously unthinned lithosphere would generate igneous intrusions with a high P-wave seismic velocity, which would thicken the crust by up to 10 km (Fig. 1c, d, blue line). Such material is not observed; rather, the oceanic crust formed first has a normal seismic velocity and is only 5.2 km thick. However, a region of high-velocity, voluminous syn-rift magmatism was found nearby in an area known as the Gop rift (Fig. 1a). Geophysical observations have shown that the Gop rift is underlain by thick oceanic crust[5] and opened before the main break-up between the Seychelles and India[20]. We find that the observed igneous thicknesses and seismic velocities can be explained if the opening of the Gop rift occurred 6 Myr before the main Deccan Traps eruption and tapped the anomalously hot asthenosphere. This initial extensional episode partly exhausted the thermal anomaly and depleted the underlying lithosphere (Fig. 1b–d, red line). Once extension had migrated southwards to the Seychelles/Laxmi ridge margin, what remained of the thermal anomaly had cooled by heat conduction. Melting was of a mildly depleted mantle that had no excess temperature, leading to thin oceanic crust seaward of the Seychelles and the Laxmi ridge with reduced seismic velocities in the lower crust.

A key question is how to reconcile these Indian Ocean results with our previous work on southeast Greenland, which suggested that previous extension focused upwelling and enhanced melt generation[18]. To answer this question, we present a more comprehensive numerical experiment that incorporates new observations from the conjugate Hatton Bank margin. This pair of margins is a similar distance from the Iceland hot spot as the Seychelles and the Laxmi ridge are from the Deccan Traps (Fig. 2a). Geological observations suggest that extension before break-up formed the Hatton Basin, thinning the continental crust by a factor of two[21]. We assign a late-Cretaceous age for this initial extensional episode on the basis of magnetic anomaly patterns[22] and the Tertiary sedimentary sequence that overlies the Hatton Basin basement rock[23]. Around 20 Myr after the formation of this basin, the earliest flood basalts were erupted on the southeast Greenland margin[9]. Interpretations

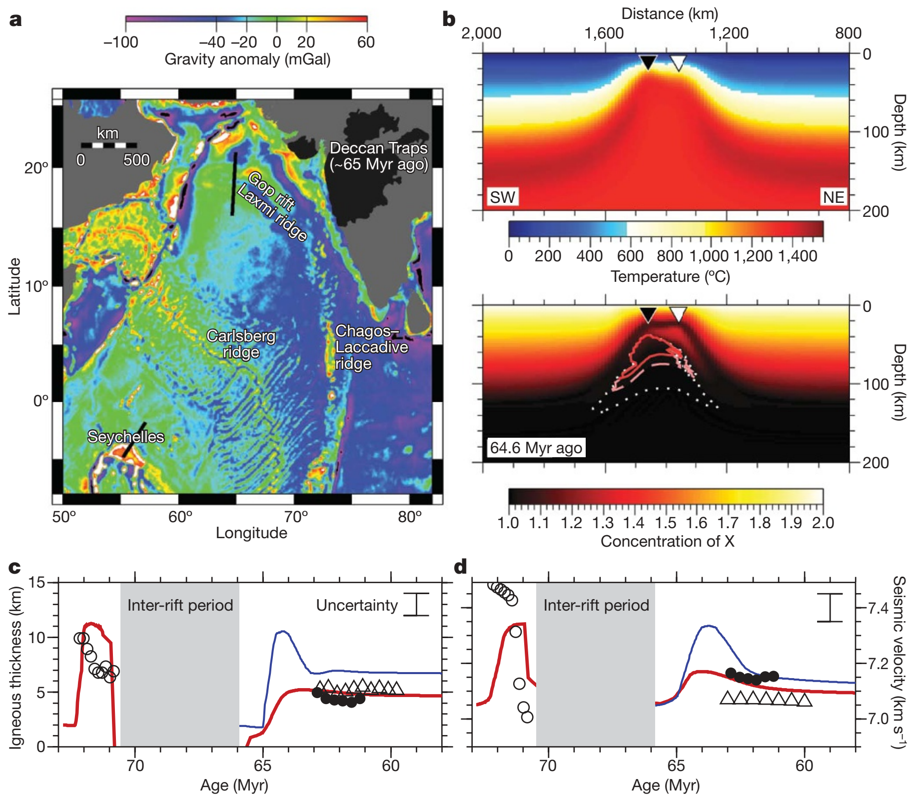

Figure 1 | Geological setting and model results for the northwest Indian Ocean. **a**, Gravity anomaly map showing the location of the conjugate Seychelles and Laxmi Ridge margin profiles. The black region marks the flood basalts of the Deccan Traps. 1 mGal = 10^{-3} cm s^{-2}. **b**, Temperature and mantle depletion (trace element X, Methods) with melt fraction at the centre of extension before break-up of the Seychelles/Laxmi ridge margin (64.6 Myr ago), from our preferred two-rift model. The location of the Gop rift is marked by the white triangle and that of the Seychelles/Laxmi ridge break-up is marked by the black triangle. In the lower plot, we show the extent of the melt region (dotted line) and the 1% and 2% melt fraction contours (dashed and solid lines, respectively). **c**, **d**, Predicted igneous crustal thickness (**c**) and predicted average lower-crustal seismic velocity (**d**) over time, assuming that the lithosphere was initially underlain by a 50-km-thick thermal anomaly at 200 °C. The blue curve is for a model in which there is a single rift event 64.5 Myr ago (half-spreading rate, 60 mm yr^{-1}) and the red curve is for a model with two rift events, one 71 Myr ago (half-spreading rate, 80 mm yr^{-1}) and the other 64.5 Myr ago[20]. In the double-rift model, the Gop rift taps the thermal anomaly and the subsequent melting in the Seychelles/Laxmi ridge margin is of a slightly depleted mantle[8]. Open circles, filled circles and open triangles respectively mark the observed igneous crustal thickness (**c**) or the average lower-crustal seismic velocity (**d**) from the Gop rift, the Laxmi ridge and the Seychelles margins[5,8]. Uncertainties in the observations are representative values taken from ref. 8.

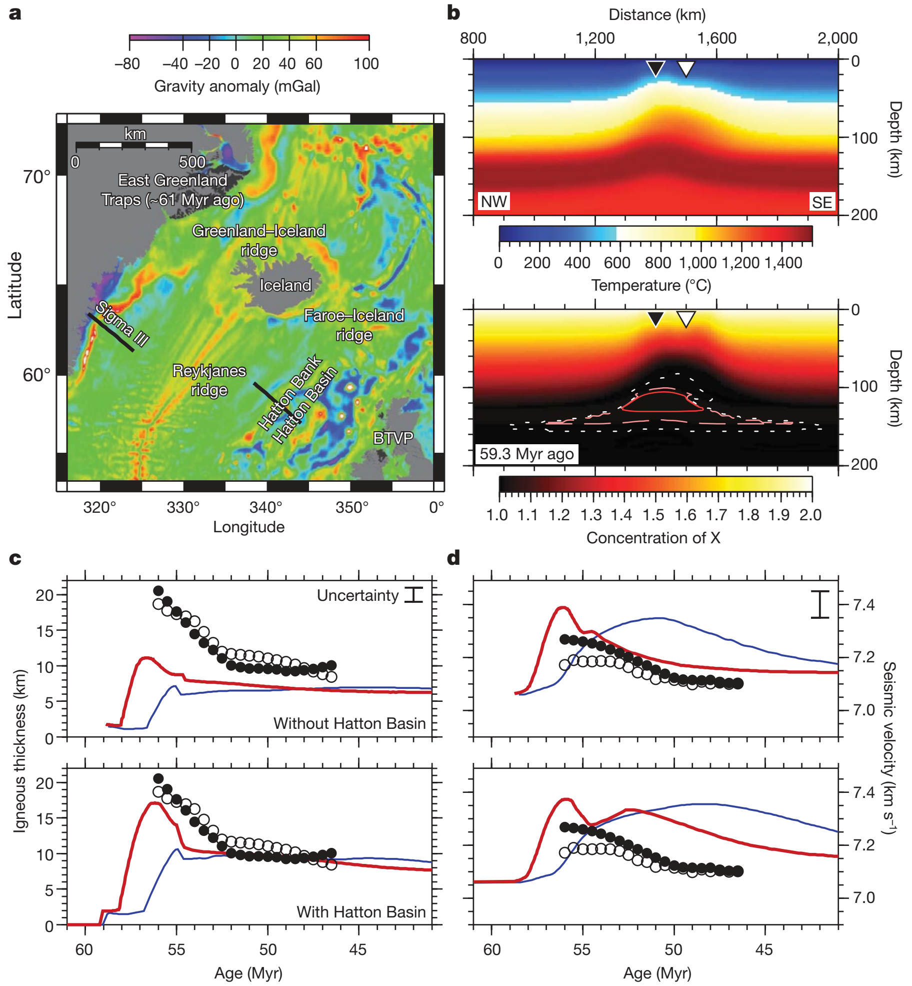

Figure 2 | Geological setting and model results for the North Atlantic Ocean. **a**, Gravity anomaly map showing the location of the conjugate southeast Greenland and Hatton Bank margin profiles[24,25]. The black regions mark the North Atlantic igneous province: the flood basalts of the East Greenland Traps and the British Tertiary volcanic province (BTVP). **b**, Temperature and mantle depletion with melt fraction, contoured as in Fig. 1, at the centre of extension before break-up (59.3 Myr ago), from our preferred model that considers the Hatton Basin opening. The location of the Hatton Basin is marked by the white triangle and that of the southeast Greenland/Hatton Bank break-up is marked by the black triangle. In the model there is a 50-km-thick thermal anomaly at 200 °C beneath the lithosphere. **c**, **d**, Predicted igneous crustal thickness (**c**) and predicted average lower-crustal seismic velocity (**d**) over time. Uncertainties in the observations are representative values taken from ref. 25. In each panel, the top plot assumes that the lithosphere is of uniform, 125-km thickness at the start of rifting between the Hatton Bank and southeast Greenland, and the bottom plot includes previous extension of the Hatton Basin that ceases 80 Myr ago. To account for competing models for the observed spreading rates, we show two alternative scenarios with half-spreading rates between Hatton Bank and Greenland that are 20 mm yr^{-1} (blue) and 40 mm yr^{-1} (red) for the first 4 Myr and then decrease to 10 mm yr^{-1} (refs 18, 25). Open and filled circles respectively mark the observed igneous crustal thickness (**c**) and the average lower-crustal seismic velocity (**d**) from the Sigma III (southeast Greenland) and iSIMM (Hatton Bank) surveys[24,25].

of initial sea-floor spreading anomalies in this part of the North Atlantic are disputed[18,24,25], and to account for these differing interpretations we show results from two models with different extension rates (Fig. 2c, d, blue and red lines). If the extension that led to the Hatton Basin is ignored, the thermal anomaly would be held beneath the 125-km-thick lithosphere. Extension in the range of the inferred half-spreading rates, although producing reasonable lower-crustal seismic velocities (Fig. 2d), would not thin the lithosphere enough to allow for significant upwelling of hot mantle to produce the 17-km-thick igneous crust observed (Fig. 2c, top). Instead, if the lithosphere is extended by a factor of two, 20 Myr before the break-up of the North Atlantic, the thermal anomaly spreads laterally beneath the slightly extended lithosphere before the eventual break-up (Fig. 2b). If the extension rate is sufficiently high, the upwelling of this hot material then generates enough melt to match the observed highly thickened igneous crust (Fig. 2c, bottom, red line) and is in reasonable agreement with the lower-crustal seismic velocity.

We conclude that the association of volcanic margins with flood basalts in the North Atlantic led to an overemphasis on the thermal structure to explain melt volumes during continental break-up. Our work shows how inherited lithosphere structure alters melting characteristics during margin formation. The interactions are individual to each margin and depend on the spatial pattern and timing of events. In the North Atlantic, extension before rifting focused upwelling and so enhanced melt generation (Fig. 3a). In the north-west Indian Ocean, previous extension exhausted the mantle thermal anomaly, leading to reduced melt generation (Fig. 3b).

Our study shows that continental rifted margins must be studied within a framework of the interaction between mantle conditions and the extension of the lithosphere. Rift history is individual to each margin, and is typically complex, but we suggest that two general considerations are required to understand melt volumes generated during break-up. First, according to our model, if thinning of the lithosphere is under way before the emplacement of a thermal anomaly, igneous crustal thickness increases by a factor of almost two across all extension rates (Fig. 4a). It is not simply the absolute lithosphere thickness that controls the melting, but the localization of thinning, or necking, of the lithosphere that focuses melt into the break-up region

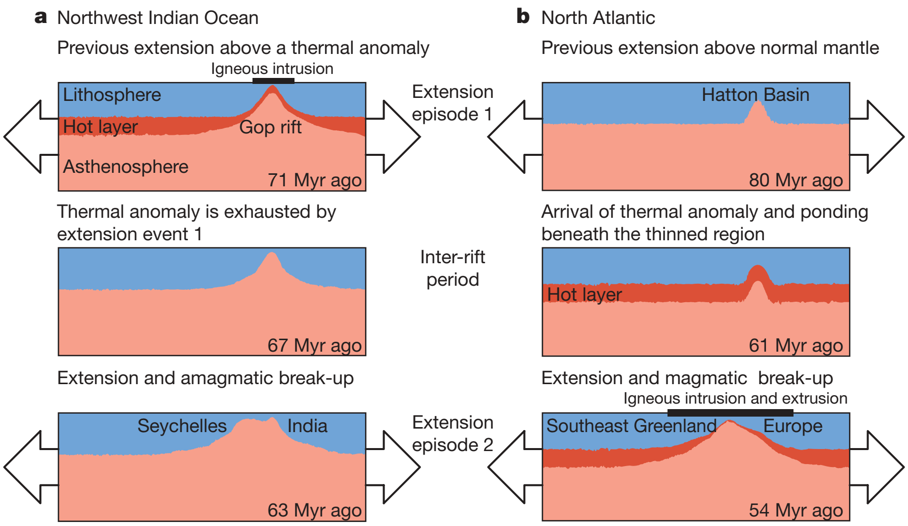

Figure 3 | Rift history controls magmatism at a rifted margin. a, Northwest Indian Ocean. The first episode of extension tapped the thermal anomaly, which is associated with the Deccan Traps. This formed the Gop rift and exhausted the thermal anomaly beneath the region of extension. The second episode of extension that led to break-up was above melt-depleted mantle and led to the amagmatic Seychelles/Laxmi ridge margin. **b,** North Atlantic. The first episode of extension formed the Hatton Bank. During the inter-rift period, the thermal anomaly ponded beneath the lithosphere, such that during the second episode of extension this thermal anomaly was tapped, leading to the high volumes of melt generated during the break-up of southeast Greenland from the Hatton Bank.

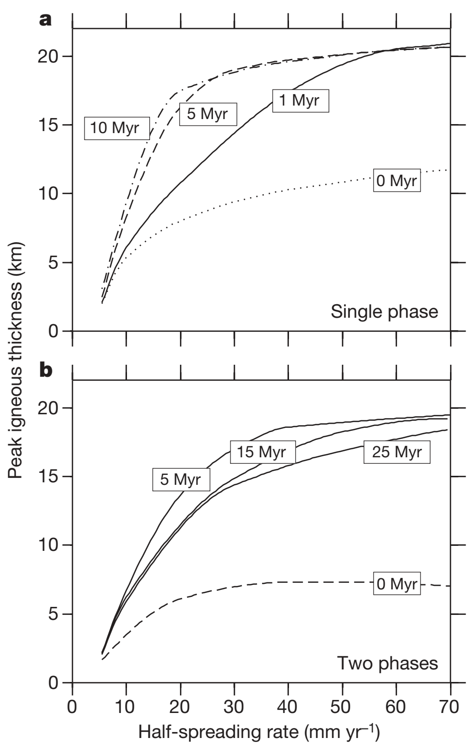

Figure 4 | The influence of rift history on melt generation during continental break-up. a, Peak igneous crustal thickness for extension at different half-spreading rates. In all model runs, break-up is achieved by a single, continuous extensional event. The four curves show cases for the arrival of a 50-km-thick thermal anomaly at 200 °C at different times after the initiation of extension. **b,** Peak break-up igneous thickness for a two-phase rift event. Extension is at the same half-spreading rate for both phases of rifting and the first extensional phase lasts for 5 Myr. There is a lateral migration of 100 km before the onset of the second extensional phase, which follows either immediately or after a time lag. The curves are labelled according to the arrival time of the hot layer (measured from the start of each model run), with the thermal anomaly arriving either at the start of phase one (dashed line) or the start of phase two (solid lines: 5 Myr, no time lag; 15 Myr, time lag = 10 Myr; 25 Myr, time lag = 20 Myr).

(Fig. 2b). This mechanism explains how, despite evidence that both the North Atlantic and northwest India mantle thermal anomalies were similar[15], the Gop rift has a maximum melt thickness of 10 km and the North Atlantic margins have more than 17 km of melt (Figs 1 and 2). Second, break-up is normally preceded by periods of extension and these can exert a strong influence on the final magmatism. Such periods of precursor extension can tap the hot mantle material, leading to relatively insignificant break-up magmatism (Fig. 4b, dashed line; cf. northwest Indian Ocean (Fig. 1)). Alternatively, if such extension took place before the arrival of a mantle thermal anomaly, it will locally thin the lithosphere. Even allowing for the effects of thermal cooling, this localized thinning enhances melt generation during break-up (Fig. 4b, solid lines; cf. North Atlantic (Fig. 2)). The degree of magmatism is critically controlled by the relative timing of the extension and the mantle thermal history. Knowledge of previous rift history is therefore required to understand the volume of magmatism observed during continental break-up.

METHODS SUMMARY

Dynamic model of extension. The lithosphere is initially 125 km thick. Mantle potential temperatures are 1300 ± 25 °C to match the range of 'normal' oceanic crustal thickness found worldwide[3,18,26]. We model anomalous mantle thermal history as a hot layer below the lithosphere. Rifting is driven by forces in the far field by imposing a divergent velocity along the top boundary. As mantle material is driven laterally by this divergence, material moves upwards to replace what has been removed. As solid mantle travels upwards it decompresses, melting when it crosses the solidus. We calculate the solidus as a function of depth and depletion. Melting depletes the solid mantle, altering the mantle composition and, hence, its subsequent melting characteristics. We track depletion by using a hypothetical completely compatible trace element, X, with an infinite partition coefficient[27]. Multiple episodes of extension are simulated by terminating the extension and restarting it in a different location. We base the timing of these events and the location of the centre of extension on geological observations. Between periods of extension, the lithosphere cools and thickens by conduction.

Calculation of observables. We estimate the average seismic velocity of the igneous crust from the major-element composition of the melt, which is calculated from empirically derived partition coefficients[28,29]. These values are then compared with the observed mean seismic velocity derived from wide-angle seismic data. To derive an estimate of the igneous crustal velocity that is unaffected by alteration or porosity, we base our mean velocities on material with a velocity greater than 6.85 km s^{-1} for all profiles[10,25]. Finally, we calculate the predicted igneous crustal thickness assuming that all the modelled melt is erupted at the ridge axis, and compare this with observations from wide-angle seismic profiles. Further details are given in the Supplementary Information.

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