

The thermal structure of volcanic passive margins

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Abstract

Over the past ten years we have numerically modelled the properties of the magmatism generated at four of the key areas where the ‘mantle plume-volcanic margin hypothesis’ is expected to be valid: The North Atlantic, South Atlantic, India-Seychelles and Afar. Our model incorporates many of the original assumptions in the classic White and McKenzie 1989 model: pure shear of the lithospheric mantle, passive upwelling and decompressional melting. Our model is however two- rather than one-dimensional, can capture the rift history (extension rate changes and axis jumps) and tracks mantle depletion during melting. In all four of our study areas we require the sub-lithospheric mantle to be 100-200 °C hotter than “normal”, non-volcanic margins to explain the characteristics of the magmatism. In the three passive margin cases we find this excess temperature is limited to a 50-100 km thick layer. We require this layer temperature to drop along-strike away from the proposed sites of plume impact at the base of the lithosphere. However, we also find that lithospheric thickness and rift history are as important as temperature for controlling the magmatism. Our work therefore lends support to the hypothesis that the excess magmatism at volcanic margins is due to a thermal anomaly in the asthenosphere, albeit with consideration of extra parameters.

Introduction

Passive margins can be classified into genetic types, such as wide or narrow, symmetric or asymmetric, non-volcanic or volcanic (e.g. Buck 1991; Brune et al. 2017; Haupert et al. 2016). Non-volcanic margins (also known as magma poor margins) are typified by a protracted duration of the extension of continental crust and the exhumation of subcontinental mantle (e.g. Whitmarsh et al. 2001). Conversely for volcanic margins, the rupture of the continent is not associated with mantle exhumation but extensive mantle melting (e.g. White and McKenzie 1989). The subdivision of non-volcanic and volcanic is not always an easy exercise, as even the archetypical non-volcanic margin may show some evidence of mantle melting (e.g. Jagoutz et al. 2007). However, the striking difference in volcanism along the global network of passive margins, for example, along the east coast of the North Atlantic from offshore Norway down to Iberia, suggests a fundamental difference in rift evolution.

In the late 1980's White and McKenzie proposed a link between continental breakup above hot asthenosphere and the formation of volcanic passive margins (White and McKenzie 1989). In their model, the magmatism is produced by decompressional melting of passively upwelling mantle during pure shear extension of the continental lithosphere. Mantle temperature was shown to be the primary control on whether a volcanic margin developed or not. They went on to show that many volcanic margins were associated with large onshore volcanic provinces and offshore aseismic ridges tracking to volcanically active islands. They therefore suggested that deep mantle plumes were the most likely source of the thermal anomaly and that these were typically 2000 km in diameter once impacted below a thinning lithosphere and raised the temperature 100–200 °C above normal.

In the following years, as more and more high-quality, deep seismic profiles were collected worldwide this model, elegant in its simplicity, started to be questioned. Examples of volcanic margins with their characteristic sequences of seaward-dipping reflectors (SDRs), high-velocity lower crustal bodies and thickened oceanic crust were found in regions that otherwise had questionable evidence for a deep mantle plume, most notably the Australian Cuvier margin (Hopper et al. 1992) and the eastern US margin (Holbrook et al. 1994a; Holbrook et al. 1994b). Alternative mechanisms to generate the degree of observed magmatism such as thermal insulation, chemical heterogeneities and small-scale convection were proposed (Mutter et al. 1988; Kelemen and Holbrook 1995; van Wijk et al. 2001; Nielsen and Hopper 2002; Korenaga 2004). Similarly, classic volcanic margins were not found in the Indian Ocean between the Seychelles and India, at a location where all other elements of the mantle plume model were present (Deccan Traps, Chagos-Laccadive-Mascarene Ridges and the active island of Reunion, Collier et al. 2009).

Establishing what causes the excessive magmatism at volcanic margins is important not only for fundamental understanding of the thermal and mechanical properties of the upper asthenosphere and lithosphere, but also has implications for resource potential. Passive margin sequences are estimated to host approximately 35% of all giant field discoveries, which in turn represent 67% of discovered conventional hydrocarbons (Levell et al. 2010). The distribution of volcanic passive margins globally is shown in Figure 1. In the Atlantic, apart from the north west Europe (Irish, UK and Norwegian sectors) and north west Africa (mainly Mauritania and Senegal) volcanic margins themselves have been relatively lightly explored for hydrocarbons. There are no commercial wells on the conjugate east Greenland margin for example and the US East Coast is under explored, both however for political rather than geological reasons (Levell et al. 2010). The volcanic margins of the South Atlantic (south of the Walvis Ridge/Rio Grande Rise) also remain relatively unexplored apart from a handful of discoveries, notably in the Orange Basin. Elsewhere the west coast of India is also unexplored whilst the north-west margin of Australia is a prolific gas province. Some of the risk associated with volcanic margins is undeniably due to problems with seismic imaging sub-syn rift volcanic targets. However concerns over source rock maturity also exist and understanding the thermal structure of these regions is critical for assessing

any prospect and it directly influences the syn and post-rift pattern of subsidence and heat-flow.

The aim of this paper is to summarise the numerical modelling work we have completed at four contrasting settings: three fully mature volcanic passive margins and one active rift. First, we will overview the numerical modelling approach taken and then outline the available observations and modelling results for each of our study areas. We will show that despite the differences in observed magmatism the four study areas present a consistent pattern of the thermal structure at the time of continental breakup. Finally we discuss the limitations of our approach and comment on the relevance of our work in a petroleum context.

A numerical model of continental breakup model

Several types of numerical models have been developed to explore the relative importance of various processes during breakup. These include complex 3-D numerical models of the viscous and plastic deformation of the upper mantle (e.g. Geoffroy et al. 2015) and 2-D viscous, elastic and plastic models of deformation (e.g. Petersen et al. 2015). However, in our work as we are mainly interested in understanding the nature of magmatism during breakup as a simplifying assumption the deformation of the crust and elastic strength of the lithosphere is ignored. This is likely justifiable, as very recent studies have found that whilst the strength of the crust will play a major role in the duration of extension and timing of volcanism, it will only have a minor effect the volume of melt generated (e.g. Pérez-Gussinye et al. 2016; Ros et al. Submitted). Therefore we focus on the role of viscous deformation in the upper mantle by exploring how the lithosphere and asthenosphere thermal and chemical structure affects the evolution of magmatism during extension.

In the appendix we describe the relatively simple self-consistent geodynamic model that includes decompression melting that we have used in our analysis. In this model extension is imposed through kinematic boundary conditions. This means we can explore the effects of mantle temperature and spreading rate on melt production, from the first syn-rift magmatism out to steady-state oceanic crust. The model was calibrated by determining the asthenospheric temperature that resulted in steady state magmatic crust thickness comparable with today's average mid-ocean ridge crustal thickness of 7 ± 1 km (Bown and White 1995). Such a thickness is achieved for the range of 1300 to 1350 °C depending on the model rate of extension (Nielsen and Hopper 2004; Armitage et al. 2008; Armitage et al. 2011). We will refer to this temperature as “baseline asthenospheric temperature” hereafter. To explore the effect of a change in the mantle temperature we either introduce a hot layer below the lithosphere (either before the start of extension or at any point during the model run) or increase the temperature of the whole asthenosphere.

Once the model is started the applied divergent velocity (extension) thins the lithosphere and causes the flow in the asthenosphere to start. This extension can stop at any time and

restart again with a laterally shifted locus. At any given time step the conditions for melting are tested, initially assuming a wet solidus, but after 2% melting a dry solidus. To compare with observations we can calculate the following properties of the igneous rocks formed by the melting (i) its thickness, which is calculated as an integral of the melt production rate assuming all melt that is generated is erupted at the centre of extension (Ito et al. 1996); (ii) its major element oxide and Rare Earth Element (REE) composition (Armitage et al. 2008; Armitage et al. 2011); and (iii) its seismic P-wave velocity, which is calculated from the major element oxide composition (Behn and Kelemen 2003). In the pre-breakup situation we can also calculate the seismic S-wave velocity of the upper mantle accounting for the effects of attenuation and melt retention (Goes et al. 2012; Armitage et al. 2015).

In practice, at each study area we first synthesised the rift history (ages, rates and locations) and estimated the initial lithospheric thickness from suitable regions adjacent to the rifted zone. We then compared the predictions (igneous rock thickness, seismic velocity and geochemistry over time) generated from models with these parameters to available observations. The igneous rock included lower crustal “underplate”, SDRs and oceanic crust, with corrections for factors such as porosity and continental rock contamination. The assessment of fit was done qualitatively by noting the general match and quantitatively by calculating misfit statistics. The input parameters of rift history and initial lithospheric thickness were also varied at each location to investigate discrepancies in the literature and to explore the sensitivity. The most significant variable tested however was the thickness and temperature of the hot layer, designed to simulate an impacted deep mantle plume at the base of the lithosphere prior to stretching and its arrival time. This extra layer contributes significantly to the syn-rift melt properties but also those of the earliest oceanic crust for all but the very slowest extension rates. Where observational data allowed, models were run at different points along the strike of a given margin.

Observations and results

The North Atlantic

Of the four study areas the North Atlantic has the most comprehensive set of geological and geophysical observations. There is a suite of modern wide-angle seismic profiles from which to measure both the thickness and seismic character of the igneous rocks generated during the rift and early drift stages (Figure 2a). The available seismic data show an approximately linear decrease in the thickness of the earliest oceanic crust from 18 km 400 km away from the inferred plume impact location (the landfall of the Greenland-Iceland-Faroes Ridge) to a ‘normal’ 7 km 1200 km away (Collier et al. 2009). This trend is observed both north and south of the aseismic ridge. The major element composition of basalts erupted on the east Greenland margin has also been measured from a series of glasses recovered during ODP leg 152 (Fitton et al. 1998). We chose therefore to model in detail a conjugate profile (SIGMA-3 – iSIMM Hatton Bank) where these observations coincided (Figure 2a). The seismic velocity

within the lower crustal intrusions on these lines is as high as ~ 7.3 km/s (e.g. Hopper et al. 2003).

In the North Atlantic breakup was only achieved after a long period of extension between the North American and European plates (e.g. Dore et al. 1999). Our modelling work shows that necking of the lithosphere during the mid-Cretaceous and the formation of extensional basins such as the Hatton Trough (see for example Chenin et al. 2015), might have been key to the extensive volcanism that was coeval with break-up during the early Cenozoic.

Within the limitations of the 2-D numerical model described in the appendix, we find that in order to recreate the observed volume and composition of melt generated offshore southeast Greenland and Hatton Bank requires two factors: (1) a mantle temperature that is 200 °C hotter than our baseline asthenospheric temperature (Armitage et al. 2008), and (2) that the continental lithosphere was thinned prior to the arrival of this mantle thermal anomaly (Figure 3; Armitage et al. 2009; Armitage et al. 2010). In the absence of increased mantle temperature, the active upwelling due to melt depletion related buoyancy is insufficient to generate large volumes of melt. In the absence of pre-thinned lithosphere, the thermal anomaly cools by heat conduction before it can influence breakup (Figure A2; Armitage et al. 2009; Armitage et al. 2010).

The South Atlantic

The evolution of volcanism in the South Atlantic margins differs from that in the North Atlantic in two important ways. (1) The volcanic margins associated with the Parana and Etendeka flood basalts lie only south of the associated aseismic ridges, the Rio Grande Rise-Walvis Ridge (Figure 1, 4b). In the North Atlantic there is an equal distribution of volcanism both north and south of the Greenland-Iceland-Faroes ridge (e.g. Collier et al. 2009). (2) There is less clear evidence for extension in the direction of final separation prior to breakup, although there may have been earlier episodes oriented north-south (e.g. Scotchman et al. 2010; Pérez-Díaz and Eagles 2014). However these events pale in comparison with the clear extension history of the North Atlantic, and are therefore unlikely to modify the thermal structure of the lithosphere regionally. There are also fewer published wide-angle profiles compared to the volcanic margins in the North Atlantic (see Becker et al. 2014), but our study has benefited from access to a comprehensive set of deep multi-channel seismic profiles from ION-GX (Figure 2b).

Reflection profiles, of course, do not give information on the P-wave velocity structure. Therefore, we took a different approach, and calculated the thickness of the earliest oceanic crust that follows breakup (Figure 3b, Taposeea et al. 2016). We subsequently compared model predictions of magmatism along-strike with the thickness of the first oceanic crustal generated (Figure 3b, central panel), as well as the reduction in oceanic crustal thickness with time (Figure 3b, bottom panel).

We find that to match the north to south reduction in initial oceanic crustal thickness, our model requires an along-strike variation in mantle temperature (Taposeea et al. 2016). The mantle needs to be 200 °C hotter than our baseline asthenospheric temperature in the north (Pelotas-Namibia), reducing to 50 °C hotter than ambient mantle in the south (Argentina- South Africa; Figure 3). Within the constraints available, the temperature drop appears approximately linear and occurs over a lateral distance of about 1800 km. This temperature range is in agreement with observations of major element oxide compositions from onshore extrusions (Trumbull 2014). Furthermore, in comparison to the North Atlantic, we believe that the thermal state of the asthenosphere is the primary control on the volcanic nature of break-up.

India-Seychelles

Our work in the Indian Ocean was motivated from our collection of the first set of modern wide-angle seismic profiles across the India-Seychelles passive margins in 2003 (Figure 2c). Given the spatial and temporal relationships with the Deccan Traps it was widely assumed that the adjacent continental margin would be volcanic (e.g. White and McKenzie 1989). However, our seismic profiles displayed surprisingly little evidence for magmatism, with limited SDRs and oceanic crust that was just 5.2 km thick (Minshull et al. 2008; Collier et al. 2009). Such thin initial oceanic crustal thickness is surprising for two reasons: First, based on studies of non-volcanic margins and mid-ocean ridges, thin oceanic crust would normally be associated with slow rates of extension (Bown and White 1994; Bown and White 1995) yet the earliest full spreading rates here are as high as 120 mm/yr (Collier et al. 2009). Second, assuming the Deccan Traps and continental break-up are related, then the mantle must have been hot or fertile (e.g. McKenzie and O'Nions 1991), and as in the Atlantic, would be expected to be associated with increased initial oceanic crustal thickness.

The new wide-angle seismic data also provided evidence for seafloor spreading in the Gop Rift, suggesting that similar to the situation in the North Atlantic i.e a rift jump was part of the separation process (Collier et al. 2008). More recent deep multi-channel seismic profiles have confirmed that the Gop Rift is most likely a short lived sea-floor spreading event which separates the continental Laxmi Ridge fragment from the Indian continent (Misra et al. 2015). Furthermore, there is strong evidence for SDRs on both margins of the Gop Rift (Misra et al. 2015). Therefore, while the eventual break-up to sea-floor spreading was magma-limited, formation of Gop Rift was associated with large amounts of volcanism.

We find that the only way to match the volume and P-wave properties of the magmatism found in the wide-angle profiles was to include the early opening of the Gop Rift (Armitage et al. 2011). Using constraints from magnetic anomaly interpretations for the timing of the two rifting events, the formation of the Gop Rift, and breakup between the Seychelles and Laxmi Ridge, we were able to produce model predictions of the magmatism that match the observations (Figure 3c). In detail, we found that volcanism within the Gop Rift is a consequence of the impingement of a thermal anomaly below the Indian continental

lithosphere, and that the rift related volcanism is genetically related to the Deccan Trap volcanism (Armitage et al. 2010). In agreement with mantle temperature inversions from the rare Earth composition of the basalts erupted within the Deccan Traps (McKenzie and O'Nions 1991), to match the volume and composition of the Gop Rift volcanism, requires an increase in mantle temperature of 200 °C above our baseline asthenospheric temperature (Armitage et al. 2011).

The Gop Rift however ultimately failed, and was most likely a short lived (<5 Myr) extensional event (Collier et al. 2008; Armitage et al. 2011). Despite its relatively short duration, we find that the extensive melting would have depleted the asthenosphere, such that when break-up finally occurred the subsequent margin was magma-limited (Figure 3c; Armitage et al. 2010). Therefore, in a similar way to our work in the North Atlantic we found that rift history was key. Here in the Indian Ocean however the relative timing of the first phase of extension, the opening of the Gop Rift, and the arrival of the hot layer worked against each other to suppress magmatism produced during the final period of breakup .

Present day breakup in Afar, Africa

The final area we have applied the same modelling approach to is Afar, Africa. The observational evidence for magmatism here differs significantly from the passive margins described above. Being dry-land it is straight-forward to determine the rock chemistry for the recent volcanism through to the eruption of the Ethiopian flood basalts at ~30 Ma (e.g. Pik et al. 2006; Rooney et al. 2012a; Ferguson et al. 2013). However, estimating the volume of material erupted is much harder as a large quantity of melt has not crystallised and likely still rests at or below the Moho (e.g. Desissa et al. 2013). However, P-to-S receiver function studies have been used to try to quantify the thickness of the crust and intrusions (Hammond et al. 2014; Reed et al. 2014). Based on these estimates, there is a thickness of between 6 and 14 km of intrusions below Afar (Armitage et al. 2015).

As in the North Atlantic and India-Seychelles margin, there is a history of extension in the Afar region. We model extension first in the southernmost Red Sea and then shift the centre of extension into the Afar depression itself. By comparing the evolution for model mantle temperatures of +50 to +150 °C against the REE composition of three volcanic massifs, we find that the thickness and composition of melt can only be matched for models at the hotter end of this range (Figure 3d). The temperature of the mantle below Afar can be further verified by converting the model temperature, pressure and melt depletion to shear wave seismic velocity and comparing magnitudes to surface wave tomography (e.g. Goes et al. 2012). By doing so, we find that the mantle below the lithosphere at Afar is likely at least 150 °C higher than our baseline asthenospheric temperature (Armitage et al. 2015).

Discussion

Alternate causes of excess volcanism during breakup

In our work we have not considered lateral variations in the chemical state of the mantle to explain volcanic margins. However, in the North Atlantic, it has been argued that the high MgO content of the volcanics could be due to heterogeneity within the mantle source rather than elevated mantle temperature. By back-calculating the primary melt composition primarily from the observations of MgO and Ni within samples collected across the North Atlantic Igneous Province, Korenaga and Kelemen 2000 suggest that there is a significant major element source heterogeneity. They then propose that the extensive volcanism observed across the North Atlantic could be due to a heterogeneity in mantle fertility (Korenaga 2004). The source of the mantle heterogeneity could be the remains of the Caledonian Orogeny that became entrained within the rifting system (Korenaga 2004; Foulger et al. 2005). Furthermore, recent thermomechanical models have demonstrated that a fossilised subduction zone of this age could indeed survive mixing in the upper mantle and influence the evolution of continental break-up (Schiffer et al. 2015).

While in certain situations it is possible that the remains of subducted slabs could influence the fertility of the asthenosphere, we believe that in the previous four sections we have demonstrated that volcanism during break-up can be comfortably explained by the thermal state of the lithosphere and asthenosphere. By taking care to model the full history of extension, the volume and composition of melt can be predicted from a relatively simple 2-D model of upper mantle deformation.

Relative importance of extension rate and rift history

Bown and White 1995 highlighted the importance of extension rate in controlling melt production during continental breakup. They showed that melt generation could be reduced by either reducing the mantle temperature or reducing the rate of extension. However, at mid-ocean ridges, which represent a simpler melting system, even ultra-slow spreading (<20 mm/yr full spreading rate) contain volcanic segments that are interspaced by regions of mantle exhumation (Carbotte et al. 2015; Cannat et al. 2009). The close proximity of volcanic and non-volcanic segments suggests extension rate is not the only controlling factor upon volcanism. Similarly, given the protracted extension observed between Iberia and Newfoundland, and the 'hyper-extended' nature of the passive margins in the South Atlantic north of the Walvis Ridge, and the South China Sea, it could be concluded that the width of a passive margin has some bearing on volcanism. However, as we have shown, ridge jumps and protracted extension can also be associated with volcanic margins. The North Atlantic has an exceptionally long history of extension, the northeast Indian margin has strong evidence of extension prior to break-up, and extension in the northernmost East African Rift migrated from the Red Sea into the Danakil Depression. Furthermore, the full rate of extension within the Danakil Depression is currently no more than 15 mm/yr (Reilinger and

McClusky 2011). Therefore it is hard to also reason that extension rate influences the volcanic nature of a rifted margin.

However, whilst we do not believe extension rate alone is a major factor, our studies have shown that the details of the rift history, in particular timing of different periods of extension is important (Armitage et al. 2010). This is because the lithosphere acts as a physical barrier to the upwards flow of buoyant hot asthenosphere (e.g. Burov and Guillou-Frottier 2005). A thicker lithosphere will reduce decompression melting because the majority of the asthenosphere will be held below the solidus. Periods of extension prior to any interaction with a thermal anomaly will therefore potentially enhance melt generation. However, the lithosphere will thermally relax in between periods of extension. This means that for the extensional history of a rift to impact melt generation, there cannot be too long a period of inactivity. For example, peak melt production, measured in terms of igneous crustal thickness, reduces by 1 km per 10 Myr of inactivity (Armitage et al. 2009).

How hot is hot?

In the presentation of the results above we deliberately referred to relative rather than absolute temperatures. This was because, for reasons touched on in the appendix, all temperature anomalies are relative to the assumed asthenosphere temperature in the model. Melt generation is calculated as a kinetic balance between the temperature above the solidus due to decompression divided by the latent heat release due to melting. We then assume all melt that is generated is erupted at the centre of extension. This calculation is highly parameterised. Consequently, we can use the observation that most of the oceanic crust is 7 km thick (Bown and White 1994), and then take the model asthenosphere temperature that generates this thickness, the range 1300 °C to 1350 °C, as representative of the normal mantle temperature.

Estimates of the present day temperature of the asthenosphere along mid-ocean ridges globally range from 1314 to 1464 °C, with the middle 50% of this distribution of measurements resting between 1355 and 1408 °C (Dalton et al. 2014). Therefore our model 'normal' asthenosphere is at least 50 °C cooler than the observed 'normal'. This is consistent with the fact that our model assumes that all melt that is generated is extracted, which will not be the case in practice. In the North Atlantic, South Atlantic and India-Seychelles the models require a mantle that was up to 200 °C hotter than normal at the time of breakup. Assuming that 1350 to 1400 °C represents normal, then we can say that the mantle below these three volcanic margins was between 1550 to 1600 °C hot at the time of formation.

In the Afar study area because of the ease of sampling volcanic rocks we are able to make a direct comparison of asthenospheric temperature. In our models we require an asthenosphere that is approximately 150 °C hotter than normal, or 1500 to 1550 °C hot. The major element oxide composition of samples indicates mantle temperatures in the range of 1370 to 1490 °C (Rooney et al. 2012b), while inverse modelling of the REE compositions

suggests a temperature of 1450 °C (Ferguson et al. 2013). Our work therefore suggests temperatures in line with the higher end of these geochemical estimates.

There are two mechanisms to deliver such hot mantle to the zone of extension: (1) convective instabilities arising from the deep mantle, plumes (e.g. Davaille and Jaupart 1993) as envisaged in the original White and McKenzie 1989 model, or (2) insulation of the mantle due to the continental lithosphere (e.g. Guillou and Jaupart 1995). It is plausible that thermal insulation can lead to an increase in mantle temperature of roughly 200 °C (Anderson 1998). However recent work shows that such a temperature increase requires the continental lithosphere to remain stable for a considerable time period of time, acting as a rigid boundary (Limare et al. 2015). The observed global variation in temperature along today's present mid-ocean ridges (Dalton et al. 2014), coupled with strong seismic evidence connecting hotter regions to the deep mantle, for example below Iceland (Rickers et al. 2013), would suggest that the simpler explanation for regions of hot mantle is that they are due to thermal plumes. In our work we also observe a decrease in asthenospheric temperature along-strike of both the North and South Atlantic continental margins, the dimension of which are consistent with current mantle plume models. It is simpler to invoke mantle plumes to explain this pattern than thermal insulation.

Implications for hydrocarbon exploration

Our work has shown a consistent mantle thermal structure of increased temperature during rifting at the three best studied volcanic passive margin locations. This result lends support to the concept of deep mantle plumes as integral components of volcanic margin development. We would therefore expect two direct consequences for the thermal history of the crust: (1) elevated mantle derived heat flow during the rifting phase and (2) increased heat delivered to the crust by the repeated intrusions (e.g. Annen and Sparks 2002). Non-conventional oil plays have been discovered around volcanic intrusions in the Neuquén Basin, Argentina (e.g. Rodriguez Monreal et al. 2009), suggesting the heat provided by the intrusion of melt into the crust alone can provide the environment in which source rocks mature.

With our geodynamic model we can explore the influence of the thermal state of the mantle on the heat flow delivered to the crust from below. Taking the preferred scenario for break-up in the southern South Atlantic (Figure 3b), we calculate the heat flow against time for the evolving passive margin (black line, Figure 4). The model predicts there to be a spike in heat flow which decays with a time scale of order 10^7 yr (10 Myr). The results obtained were especially sensitive to the initial lithospheric thickness, if it is increased or decreased by 40 km the peak heat flow changes by up to 50%, although the overall trend remains unchanged.

Our model is unable to directly estimate the heat flow resulting from the second factor, namely the addition of heat by crustal intrusions. The model does not consider what happens to the melt produced, whether it intrudes the lower continental crust, inject as sills

within the sediments or erupt as flood basalts for example. Clearly, where the hot material is emplaced will affect the thermal structure on a local scale (White et al. 2003; Fjeldskaar et al. 2008). However, we can draw some general conclusions based on our modelling work.

At volcanic margins the observed thickness of the syn-rift magmatism varies between 10-20 km. Our forward models predict that such magmatism is generated within a 5-10 Myr period (Figure 3). For a 10 km addition of magma over 5 Ma this would require repetitive intrusions at the scale of 20 m at a periodicity of 10^4 yr. Annen and Sparks 2002 showed that repetitive intrusions of this order of magnitude would increase crustal temperatures to 800 °C, and that the decay of this crustal anomaly once volcanism ceases would take a few million years. Note however that repeated intrusions of 50 m with a period of 10^5 yr will have a significantly smaller impact on the crustal heat flow (Annen and Sparks 2002).

The thermal decay timescale of an intrusion can also more roughly be estimated from the thermal diffusivity of the crust, $\kappa = k/(\rho C_p)$, where $k \sim 3.0 \text{ Js}^{-1}\text{m}^{-1}\text{K}^{-1}$ is the thermal conductivity of the crust, $\rho \sim 2650 \text{ kgm}^{-3}$ is the crustal density, and $C_p \sim 1370 \text{ Jkg}^{-1}$ is the specific heat capacity. The time scale for the decay of the heat anomaly generated by the intrusion of melt can then be estimated from the diffusion timescale, L^2/κ , where L is the thickness of the intrusion. Assuming a 10 km thick intrusion the decay time scale is of the order of 10^6 yr, in agreement with the more complete analysis of Annen and Sparks 2002 and for example the 2-D stratigraphic model of Rodriguez Monreal et al. 2009.

Therefore in volcanic margin settings there are two sources of heat flow: (1) that which is due to the history of stretching of the continental lithosphere and decays with a time scale of order 10^7 yr (Figure 4); and (2) the heating introduced by the intrusion of melt, that has a more rapid decay. Two final considerations is that any magmatism generated during continental breakup also adds a volume to the continental, crust which should not be ignored during subsidence modelling (White et al. 2010), and may act as a barrier to fluid migration (e.g. Holford et al. 2013).

Conclusions

Our work has shown a consistent mantle thermal structure at the time of breakup at the three best studied volcanic passive margin locations, and lends support to the concept of deep mantle plumes as integral components of volcanic margin development. In all four of our study areas we require the sub-lithospheric mantle to have an excess temperature of 100-200 °C to explain the characteristics of the magmatism. In the three passive margin cases we find this excess temperature is limited to a 50-100 km thick layer. We also require this layer temperature to drop along-strike away from the proposed sites of plume impact at the base of the lithosphere. The thermal anomaly is in all cases responsible for the excess volcanism observed, however the maximum magnitude of volcanism is enhanced if the lithosphere is thinned prior to break-up. From a thermal modelling perspective we suggest there are two time scales of interest, the long term, order 10^7 yr decay, of the mantle heat

flow anomaly related to stretching and a shorter 10^6 yr decay timescale for crustal heating due to the intrusion of melt. This second rapid heat and cooling will be sensitive to melt generation and hence the thermal structure of the mantle at the time of break-up.

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Figures

1) Map of major continental margin types (based on Reston 2009; Haupert et al. 2016 and others). Continental shelves (dark grey shading) and plate boundaries (fine lines) are from Muller et al. 2008. Black shading are Large Igneous Provinces (LIPS), including continental flood basalt provinces, oceanic plateaus, submarine ridges, ocean basin flood basalts and seamount groups but excluding volcanic continental margin for clarity (Coffin and Eldholm 1994). Yellow stars mark the areas we have studied. Proposed associated modern day plumes are labelled IC Iceland; TC Tristan da Cunha; RE Reunion and AF Afar.

2) Maps of the four study areas showing the locations of the observational data used. (a) North Atlantic showing the East Greenland Trap-British Tertiary Volcanic province (BTVP) continental flood basalt provinces, Greenland-Faroes aseismic ridge and modern active volcanic island of Iceland. The bold black lines show wide-angle seismic lines included in the compilation of early oceanic crustal thickness given in Collier et al. 2009 and the red lines show the conjugate wide-angle profiles modelled in detail in Armitage et al. 2010 and shown in Figure 3a. (b) South Atlantic showing the Parana-Etendeka continental flood basalt provinces, Walvis-Rio Grande aseismic ridge and modern active volcanic island of Tristan da Cunha. The bold black lines show wide-angle and green lines deep multi-channel seismic lines respectively included in our analysis of the early oceanic crustal thickness given in Taposeea et al. 2016. Red lines mark the profiles included in Figure 3b. (c) Indian Ocean showing the Deccan Traps continental flood basalt provinces, Chagos-Laccadive aseismic ridge and modern active volcanic island of Reunion (to south of area shown). The red lines show the conjugate wide-angle profiles modelled in detail in Armitage et al. 2010 and shown in Figure 3c. (d) Afar. The magmatic crustal thickness and geochemistry measurements shown in Figure 3d come from locations throughout the Afar triangle (see Armitage et al. 2015 for details).

3) Key model results for the four study areas. The top figures show the thermal conditions within the model at the time step shown by the pink bar in the lower diagrams. The black outlines show the regions that are producing melt with contours representing 1% (long dashed lines) and 2% (short dashed lines) melting. The lower figures show examples of observations (symbols or in the Afar case a grey box) compared to model predictions (red

lines). We have chosen to show models from early stage of rifting (North and South Atlantic) and from late stage (South Atlantic and Afar). In the former two examples there are rift jumps, hence the melt area shown is asymmetric. There is no other significance to the choice of time step for each location. (a) North Atlantic: the lithosphere has been pre-stretched by the formation of the Hatton Trough at 5 mm/yr (see Armitage et al. 2009), and subsequently the event that leads to break-up at a half spreading rate that is initially 40 mm/yr and reduces to 10 mm/yr at 55 Ma. The model is compared to estimates of the thickness and Vp of igneous intrusions from the SIGMA-3 – iSIMM Hatton Bank conjugate margins (Hopper et al. 2003; White et al. 2008). (b) South Atlantic: the lithosphere has been stretched in one phase at a half spreading rate of 14 mm/yr (see Taposeea et al. 2016). The model is compared to the thickness of the oceanic crust off-shore Namibia (Mamba line) and Pelotas (ION-GX). (c) The India – Seychelles: the lithosphere has been stretched in two phases, first the formation of the Gop Rift at a half spreading rate of 80 mm/yr, followed by extension between the Seychelles and Laxmi Ridge at a half spreading rate of 60 mm/yr (see Armitage et al. 2011). The model is compared to estimates of the thickness and Vp of igneous intrusions from both margins (Minshull et al. 2008; Collier et al. 2009). (d) Afar, Africa: the lithosphere has been stretched in two phases, first due to the formation of the southernmost Red Sea at 5 mm/yr, and then within Afar at a half spreading rate of 7 mm/yr (see Armitage et al. 2015). The model is compared the estimates of the thickness of the igneous crust and the Dy/Yb ratio of melts erupted.

4) Predicted mantle heat flow at the margin of the evolving rift zone. The heat flow is calculated from the geothermal gradient at the model surface at a point that tracks the margin as the model evolves. Initially this point is in the centre of the extending basin, but once melt is generated it migrates with the model such that it maintains its position at the edge of the continent/transitional crustal boundary (see green arrow in the cartoons on the right). The base model is the same as used to predict the volume of melt erupted in the southern South Atlantic (Figure 3b) with a 200 °C hot 100 km thick layer below. Half spreading rate is 12 mm/yr. Three cases are shown with different initial continental lithosphere thickness. The cartoons on the right illustrate the stages leading up to breakup and beyond for the 180 km thick initial lithosphere model (red line in the graph). Melt starts to be generated after 3, 8 and 13 Myr for the 100, 140 and 180 km thick initial continental lithospheric thickness respectively with breakup (defined here as when the continental lithosphere is essentially thinned to zero such that all the melt produced generates oceanic crust) occurs at 10.0, 14.2 and 17.6 Myr respectively.

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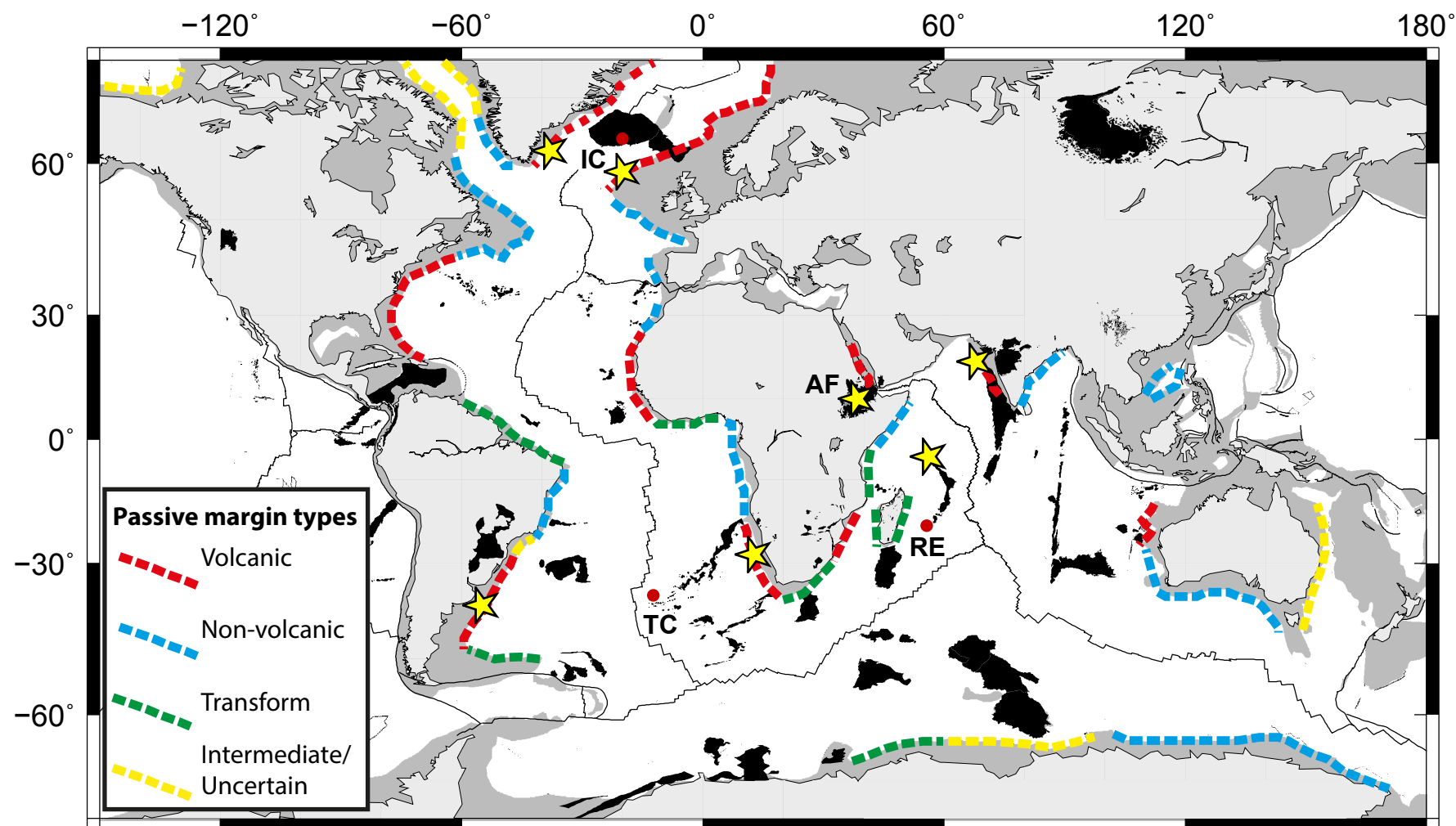
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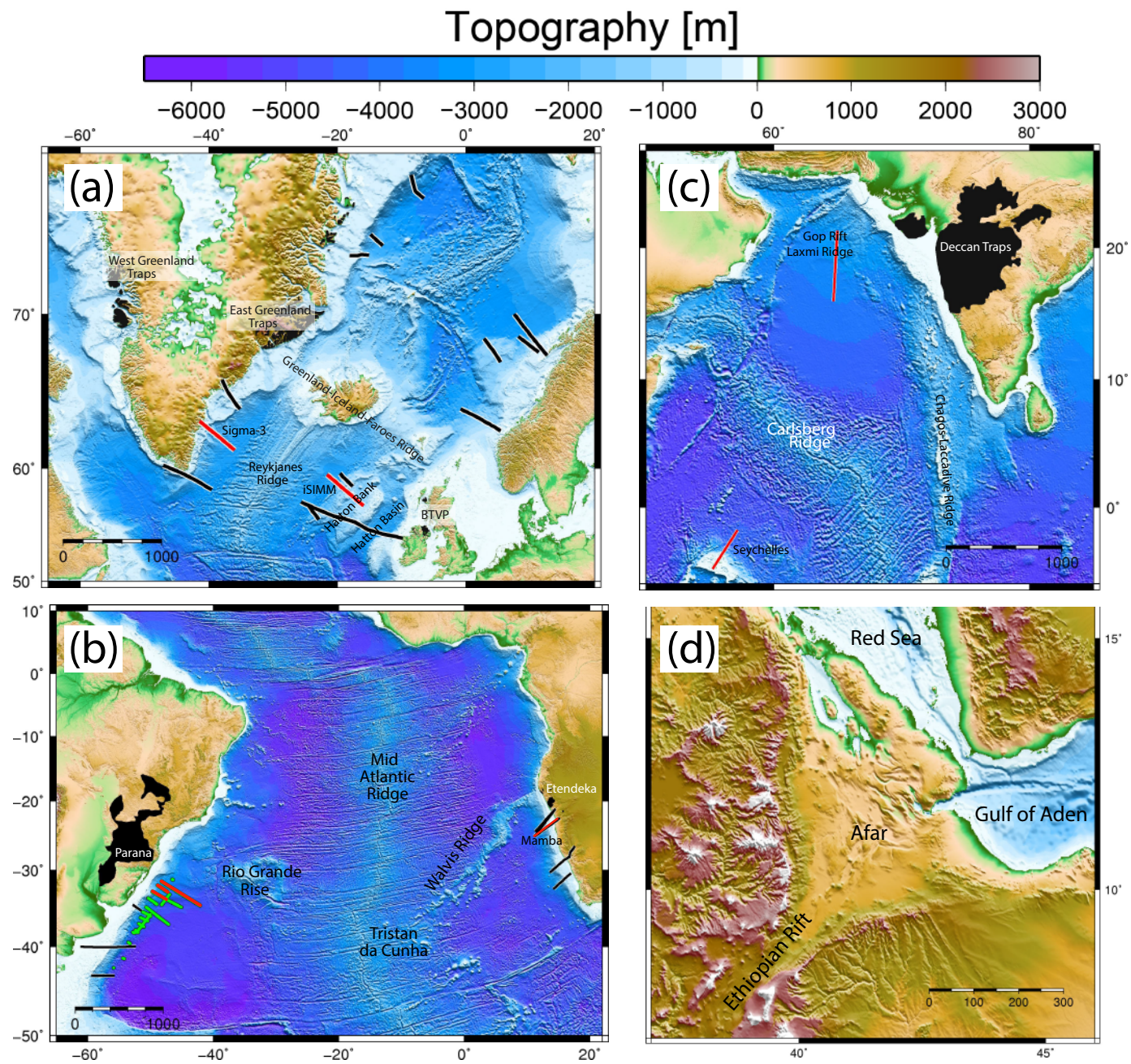
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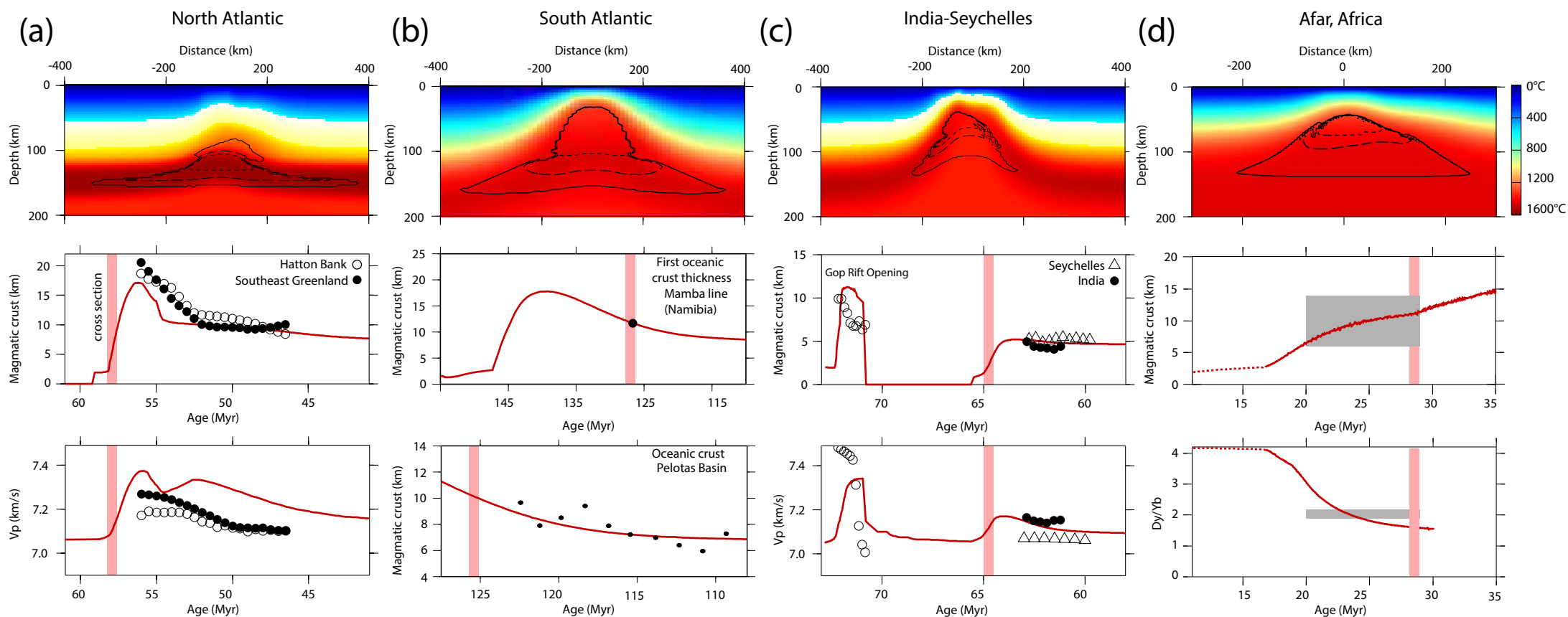
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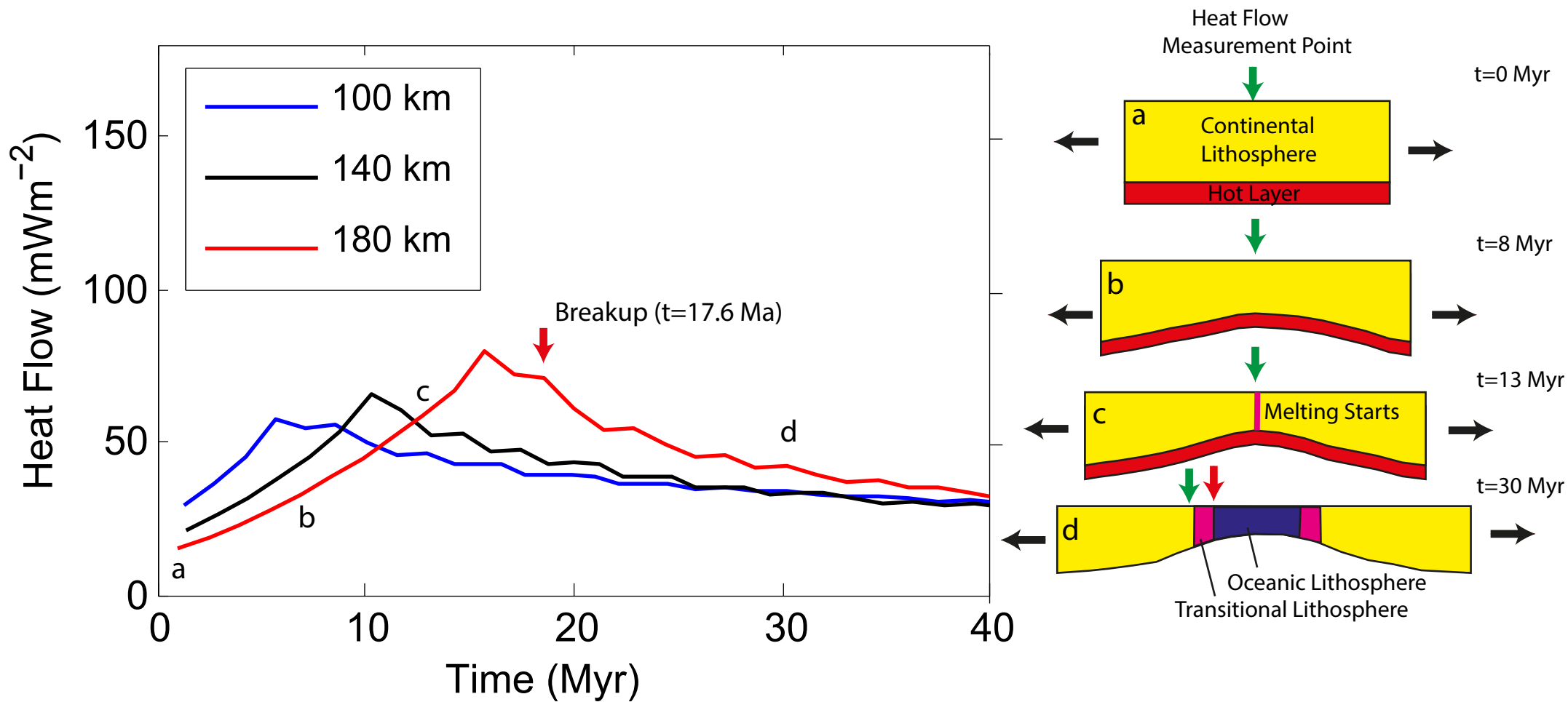
Armitage & Collier Fig1



Armitage & Collier Fig2



Armitage & Collier Fig3

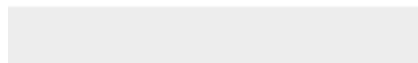
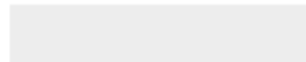


Armitage & Collier Fig4



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