A review of Pangaea dispersal and Large Igneous Provinces – in search of a causative mechanism

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Abstract

The breakup of Pangaea was accompanied by extensive, episodic, magmatic activity. Several Large Igneous Provinces (LIPs) formed, such as the Central Atlantic Magmatic Province (CAMP) and the North Atlantic Igneous Province (NAIP). Here, we review the chronology of Pangaea breakup and related large-scale magmatism. We review the Triassic formation of the Central Atlantic Ocean, the breakup between East and West Gondwana in the Middle Jurassic, the Early Cretaceous opening of the South Atlantic, the Cretaceous separation of India from Antarctica, and finally the formation of the North Atlantic in the Mesozoic-Cenozoic. We demonstrate that throughout the dispersal of Pangaea, major volcanism typically occurs distal from the locus of rift initiation and initial oceanic crust accretion. There is no location where extension propagates away from a newly formed LIP. Instead, LIPs are coincident with major lithosphere-scale shear movements, aborted rifts and splinters of continental crust rifted far out into the oceanic domain. These observations suggest that a fundamental reappraisal of the causes and consequences of Gondwana-breakup-related LIPs is in order.

1.0 Introduction

Throughout geological time the majority of continental lithosphere has several times been assembled into supercontinents (Rogers, 1996; Stampfli et al., 2013; Frizon De Lamotte et al., 2015; Merdith et al., 2019) (Fig. 1). The processes that initiate the dispersal of these large continental accumulations remain controversial (Santosh et al., 2009; Audet and Bürgmann, 2011; Murphy and Nance, 2013; Nance et al., 2014; Petersen and Schiffer, 2016; Peace et al., 2017a; Petersen et al., 2018; Schiffer et al., 2018; Olierook et al., 2019). The debate primarily revolves around whether continental dispersal is driven by deep-rooted thermal anomalies (Morgan-type mantle plumes) or shallow plate tectonic processes (Storey, 1995; Dalziel et al., 2000; Beutel et al., 2005; Frizon De Lamotte et al., 2015; Pirajno and Santosh, 2015; Yeh and Shellnutt, 2016; Keppie, 2016; Petersen et al., 2018; Heron, 2018).

The concept that plumes from the deep mantle are the main driver of continental rifting was originally proposed by Morgan (1971) who suggested plumes provide "*the motive force for continental drift*" and that "*currents in the asthenosphere spreading radially away from each upwelling will produce stresses on the bottoms of the lithospheric plates which, together with the stresses generated by the plate to plate interactions at rises, faults and trenches, will determine the direction in which each plate moves*". Despite the fact that continental breakup can often be magma-poor (Whitmarsh et al., 2001; Reston, 2009; Franke, 2013) this hypothesis continues to be commonly invoked as a default to explain continental breakup and plate motions, particularly the case where rifting is accompanied by major magmatism (Richards et al., 1989; White, 1992; Campbell and Kerr, 2007).

Alternative models have nevertheless been proposed. The coincidence between the primary Atlantic "hot spots" and the spreading plate boundary has been pointed out (Julian et al., 2015), as has their persistence in near-ridge localities. Such a causal relationship means that those "hot spots" cannot be stationary relative to the underlying mantle. That observation has inspired a number of models including ones that attribute the excess volcanism to fusibility in the source, brought about by excess volatiles (e.g., Bonath, 1990; Ligi et al., 2005) or enhanced source fertility resulting from recycled near-surface materials (Foulger and Anderson, 2005; Foulger et al., 2005b). In the plume model, the persistence of the excess volcanism on the ridge is attributed to "upside-down drainage", i.e., lateral flow of hot material from a non-ridge-centred, migrating plume, along the underside of the lithosphere to an eruptive site where the lithosphere is thinnest (Sleep, 1996).

A number of non-ridge-centred "hot spots" have also been proposed to lie in the Atlantic, including at Bermuda, the Canary Islands, the Cape Verde Islands and St. Helena. It remains an unanswered question why, if they are fed by deep-mantle plumes, are their products not also channelled to the spreading ridge. Some of these have been attributed to different plume- and non-plume origins, often based on the geochemistry of their lavas. This may be variable, in particular in the source water contents. Proposed mechanisms include lateral flow from a nearby, branching mantle plume (e.g., the proposed multi-headed Tristan plume), flexure of the edge of the continental shelf resulting in lithosphere rupture (e.g., for the Canary and Cape Verde volcanism), and extraction of melt from the low-velocity zone that is ubiquitous beneath the lithosphere (e.g., Presnall and Gudfinnsson, 2011). Shear heating resulting from motion of the plates must also contribute heat and induce formation of partial melt in the asthenosphere and can account for the availability of melt away from plate boundaries everywhere (Doglioni et al., 2005).

To date, there has been limited discussion of whether the rifting process in itself can account for the excess volcanism observed (Peace et al., 2017a). This is likely partly because there has been relatively little attention paid to modelling the volumes, and ranges in volume, of melt observed (Petersen et al., 2018). Where this has been done, the results are compelling. Asthenospheric upwelling is an inevitable consequence of lithospheric rifting, regardless of the driving mechanism (Huismans et al., 2001; Merle, 2011). In addition, numerical models that include small-scale upwelling can reproduce LIP-scale volumes of melt (Simon et al., 2009).

However, not all rifted margins contain large amounts of magma and there is a continuous spectrum between ‘magma-rich’ and ‘magma-poor’. This is popularly attributed to the presence or absence of a nearby mantle plume or thermal anomaly, thereby attributing it to variations in temperature of the source. Drawing from the alternative, non-thermal models that have been proposed for on-ridge excess magmatism, an explanation in variations in source fusibility and fertility also presents a feasible explanation (Korenaga and Kelemen, 2000; White et al., 2003; Korenaga, 2004; Petersen and Schiffer, 2016; Peace et al., 2017b).

The great structural diversity of continental rifts testifies to their dependence on not just one but many factors (Şengör and Natal’in, 2001; Merle, 2011). Rifts develop in different tectonic environments, on diverse pre-existing structures (Doré et al., 1997; Petersen and Schiffer, 2016; Schiffer et al., this volume), and under slow- or fast-extending conditions (Lundin et al., 2018). They may evolve to form narrow or wide extending zones (Davison, 1997), be magma-poor or magma-rich (Franke, 2013), and exhibit asymmetric or symmetric extension (Becker et al., 2014; Peace et al., 2016).

In this article we review the spatial and chronological relationships between large-volume magmatism and rifting to synthesise the large volume of material already published on this topic rather than introduce new analyses. We analyse in detail Pangaea’s dispersal (Fig. 1) in relation to LIPs and other magmatism to test the predictions of a causal relationship between proposed plumes and continental rifting in the Pangaean realm. Specifically we test two predictions of the active rifting hypothesis, one chronological and the other kinematic:

1. Large-scale volcanism is generated during or just after lithospheric doming but before rifting and breakup (chronological); and

2. Rifting and breakup initiates at the location of thermal uplift and propagates away from it (kinematics).

2.0 The assembly and dispersal of Pangaea

Pangaea was constructed from multiple lithospheric plates that resulted from the disintegration of the previous supercontinent Rodinia. Before Rodinia broke apart in the Late Proterozoic, between 1000 and 700 Ma (Veevers, 2004), it comprised North America, Baltica, Siberia, Gondwana, and other minor components (Torsvik et al., 1996; Stampfli et al., 2013). The disassembly of Rodinia is poorly understood, as geological evidence has been overprinted by later orogenic cycles (Scotese, 2009; Li et al., 2008; Cawood and Pisarevsky, 2006), while the assembly and disassembly of Pangaea is better understood and captured in detailed palaeogeographic reconstructions (Golonka et al., 1994; Stampfli et al., 2013; Blakey and Wong, 2003; Cocks and Torsvik, 2006; Scotese, 2009) (Fig. 1).

Pangaea's earliest breakup and formation of the first oceanic crust occurred in the Triassic and formed the Central Atlantic Ocean. Subsequently, West Gondwana (Africa and South America) and East Gondwana (Antarctica, Australia, India, Madagascar, and New Zealand) started to separate in the Middle Jurassic. This was followed by the Early Cretaceous separation of Africa from South America during the opening of the South Atlantic, then the Cretaceous separation of India from Antarctica, and finally the successful breakup of Scandinavia and Greenland, and the birth of the North Atlantic in the Cretaceous to Early Cenozoic (Fig. 1). Here, we investigate if volcanism associated with each breakup event occurred before, during, or after rifting. We also review the kinematic evolution of each rift and their initial breakup positions in relation to LIPs throughout the dispersal of Gondwana and Laurasia.

The Carboniferous–Permian assembly of Pangaea was preceded by four major tectonic events:

1) Disassembly of the supercontinent Rodinia in the late Proterozoic, when Laurentia, Baltica, and Siberia separated from a number of other continents, opening the Iapetus Ocean and the Tornquist Sea between them. Gondwana formed shortly afterwards, in the Cambrian, by re-assembly of the remaining dispersed continents of India, Australia, Sahara, West Africa and other minor blocks (Li et al., 2008).

2) In Ordovician-Devonian times, the Caledonian Orogeny sutured Laurussia (North America, Baltica and Avalonia) which had previously drifted northward from Gondwana forming the Rheic Ocean (Cocks and Torsvik, 2011).

3) In the Devonian, peri-Gondwana terranes, rifted from Gondwana, opened the Palaeotethys and accreted to southern Laurasia during the Devonian Variscan Orogeny. Siberia and Kazakhstan docked along the eastern Laurussian margin, during the Uralide Orogeny to form Laurasia. This was followed by Carboniferous-Jurassic collision between Gondwana and Laurasia along the Appalachian fold belt, finally assembling western Pangaea (Stampfli and Borel, 2002; Cocks and Torsvik, 2007).

4) Assembly of eastern Pangaea (central and SE Asia) in the late Permian-Jurassic involved the closure of the Palaeotethys and the Mongol–Okhotsk Ocean to accrete peri-Gondwana terranes to Siberia and Kazakstan in the Late Jurassic (Zorin, 1999; Kravchinsky et al., 2002; Sengor, 1996; Tomurtogoo et al., 2005). The repeated rifting of peri-Gondwana terranes opened the Neotethys.

The dispersal of Pangaea (Fig. 1) occurred through an extended period of Earth's history and is well-summarised in earlier papers (e.g., Dietz and Holden, 1970; Frizon De Lamotte et al., 2015).

Rifting began in western Pangaea in the Triassic-early Jurassic, coeval with the final phases of the assembly of eastern Pangaea, initiating the disassembly of Pangaea. In the mid-Jurassic, continental breakup and seafloor spreading opened the Central Atlantic-Caribbean (Biari et al., 2017) and the Indian Ocean (Powell et al., 1988), breaking Pangaea apart again between North America, West Gondwana (South America and Africa) and East Gondwana (India, Antarctica, Madagascar and Australia) (Schettino and Scotese, 2005). By the end of the Early Cretaceous, East Gondwana was completely detached from West Gondwana, while India separated from Antarctica and Australia and the Amerasia Basin opened in the Arctic. Rifting leading to seafloor spreading started separating South America and Africa from south to north, finally adjoining with Central Atlantic Ocean spreading in the mid-late Cretaceous. Madagascar began diverging from Africa in the Middle Jurassic (Phethean et al., 2016). This was followed by the Labrador Sea opening in the Late Cretaceous (Roest and Srivastava, 1989; Roest and Srivastava, 1989; Chalmers and Pulvertaft, 2001; Peace et al., 2016; Peace et al., 2018a; Abdelmalak et al., 2018), along with the Gulf of Aden (Courtillot, 1980). Early in the Cenozoic, the Labrador Sea was gradually abandoned in favour of rifting between North America-Greenland and Europe which opened the NE Atlantic in the Palaeocene (Srivastava, 1978; Gaina et al., 2017b). The opening of the North Atlantic represents the dispersal and end of the Laurasian continental amalgamation that formed the northern constituent of the Pangaea supercontinent (Hansen et al., 2009; Gaina et al., 2009; Frizon De Lamotte et al., 2015). At the same time Australia separated from Antarctica and Zealandia (Veevers, 2012; Williams et al., 2019), and the Gakkel Ridge started opening the Eurasia Basin in the Arctic (Thórarinsson et al., 2015).

3.0 The opening of the Central Atlantic

The Central Atlantic is defined here as the region bounded to the north by the Pico and Gloria fracture zones and to the south by the Fifteen-Twenty and Guinean fracture zones (Fig. 2). This oceanic basin comprises the oldest part of the Atlantic Ocean, with oceanic crust dating back to the Triassic-Jurassic boundary (Biari et al., 2017). As this region contains the earliest breakup and formation of oceanic crust, it is a prime region for understanding the whole Atlantic system, including the North and South Atlantic Oceans. Conversely, this area is difficult to explore due to the many complexities involved in the rifting process (Pindell and Dewey, 1982; Reston, 2009).

The North American-African segment of the Central Atlantic has undergone multiple suturing and breakup events along similar axes over at least two Wilson Cycles, suggesting a major control of inheritance in this region (Wilson, 1966; Pique and Laville, 1996; Thomas, 2018). Furthermore, the continental margins are buried below voluminous salt bodies, making seismic imaging difficult (e.g., Labails et al., 2010). In addition, dating oceanic crust older than Chron M-25 (~155 Ma) has proven problematic because of the Jurassic magnetic quiet zone (Roeser et al., 2002). Breakup of the Central Atlantic was contemporaneous with significant magmatism, namely the Central Atlantic Magmatic Province (CAMP), one of the most significant LIPs which may correspond to the end-Triassic mass extinction (Marzoli et al., 1999; Verati et al., 2007; Nomade et al., 2007; Panfili et al., 2019).

3.1 Overview of Central Atlantic rifting and breakup

The Central Atlantic Ocean opened after a protracted period of rifting (Biari et al., 2017), which led to the formation of major rift basins on the continental margins (Withjack et al., 2012), and is claimed to have displayed significant asymmetry between the Scotian and the Moroccan margins (Maillard et al., 2006). Several ridge-jumps may have occurred during early opening (e.g., Labails et al., 2010). There is also a significant difference in rifting style between the northern and southern parts (Leleu et al., 2016). Extension began in the northern Central Atlantic in the Anisian (Middle Triassic) and the Carnian (Late Triassic) in the southern Central Atlantic, long-lived passive rifting preceded emplacement of the Central Atlantic Magmatic Province) CAMP at ~201 Ma (Leleu et al., 2016).

Seafloor spreading is thought to have started around 180–200 Ma, either during the Late Sinemurian (195 Ma) (Sahabi et al., 2004) or the Middle Jurassic (175 Ma) (Klitgord and Schouten, 1986). Labails et al. (2010) suggested that the opening of the Central Atlantic started during late Sinemurian (190 Ma), and that initial spreading (up to 170 Ma) was characterised by extremely slow crustal production (~0.8 cm/y half spreading rate). In addition, Labails et al. (2010) show that at the time of the Blake Spur Magnetic Anomaly (BSMA) (170 Ma), the direction of the relative plate motion between Laurentia and Africa changed from NNW-SSE to NW-SE and the half spreading rate increased to ~1.7 cm/y. Labails et al. (2010) also identified a conjugate magnetic anomaly to the BSMA, which they suggest rules out the possibility of a ridge jump. Labails et al. (2010) further reports a significant spreading asymmetry, producing more oceanic crust on the American plate. In addition to the temporal variation in spreading rates identified by Labails et al. (2010), spreading rates in the northern Central Atlantic are thought to be lower than those of the southern Central Atlantic (Klitgord and Schouten, 1986).

While many existing plate reconstructions show isochronous breakup along the whole margin, a detailed analysis of tectonic structures shows differences in the timing for the American margin (Withjack et al., 1998). In particular, Withjack et al (1998) showed that the rift-drift transition offshore of the SE USA took place at around 200 Ma, while offshore Canada this transition is dated to around 185 Ma. Le Roy and Piqué (2001) analysed rift structures at the Moroccan margin and found a westward migration of extension during Carnian to Rhaetian (Late Triassic) times. They conclude that oceanic accretion could have already started in the early Lower Jurassic. With the assumption of a half spreading rate of 0.8 cm/y (Labails et al., 2010), an interpolation of magnetic anomalies by Roeser et al. (2002) yielded an age estimate for the initial ocean crust offshore Morocco of 193.5 Ma. By forward modelling of magnetic measurements, Davis et al. (2018) concluded that the formation of Seaward dipping reflector (SDR) packages most probably has taken place at a relatively low extension rate (< 2 cm/y full-spreading). The width of the SDRs suggests that formation of a complete SDR wedge would have taken at least 6 Myr. Assuming that the emplacement of SDRs started directly after the emplacement of the CAMP LIP, Davis et al. (2018) concluded that the earliest oceanic crust within the Central Atlantic has an age of ~195 Ma or younger.

3.2 Rifting and magmatism

The opening of the Central Atlantic was contemporaneous with the production of extensive dykes, sills, and surface flows along the margins and interiors of eastern North America, NE

South America, NW Africa, and southwestern Europe (Hodych and Hayatsu, 1980; Papezik and Hodych, 1980; Deckart et al., 2005; Nomade et al., 2007; Kontak, 2008; Bensalah et al., 2011; Shellnutt et al., 2017; Denyszyn et al., 2018). This association of basaltic magmatism with continental rifting and breakup indicates features and mechanics of the mantle during both events. The CAMP is certainly one of the largest and most important LIPs globally recognised (Bryan and Ernst, 2008).

Since the 1970s, similarities between Early Mesozoic basalts on the margins of eastern North America and NW Africa have been recognised (e.g., Weigand and Ragland, 1970; May 1971; Bertrand and Coffrant, 1977). The term "CAMP" was first used by Marzoli et al. (1999), who included dykes and sills in NE South America. The extent of the CAMP is primarily defined in previous work by the location of dykes, with the CAMP boundaries drawn based on their farthest known extent. The petrology of the igneous rocks comprising the CAMP distinguishes them from the older and younger basaltic intrusions in the same regions (e.g., Merle et al., 2013). Swarms of related dykes tend to occur in distinct sets of dozens to hundreds with similar orientations and field characteristics. Sills of the CAMP occur both within Mesozoic basins and also in older crustal rocks in South America and Africa. Large tholeiite sills are also mapped in Brazil and western Africa (Davies et al., 2017; Marzoli et al., 1999), while smaller but still-considerable examples are well known in the eastern USA in the Hartford, Newark, and Deep River Mesozoic basins, though not in the older basement rocks.

Mesozoic basins that preserve CAMP extrusive basalts cover a total area of about 300,000 km^2 (McHone, 2003). However, dykes and sills of the CAMP that fed the basin basalts also occur across 11,000,000 km^2 within four continents, centred upon but extending far outside of the initial Pangaean rift zone (Fig. 2). The breadth of the CAMP exceeds 5,000 km, with several dykes longer than 500 km, sills exceeding 100,000 km^3, and lava flows possibly larger than 50,000 km^3 (McHone, 1996). If only half of the continental CAMP area was originally covered by 200 m of lava, the total volume of the CAMP and the East Coast Margin Igneous Province (ECMIP; the thick rift-related igneous package interpreted to underlie the North American Central Atlantic margin e.g., Holbrook and Kelemen, 1993) extrusive basalt would exceed 2,400,000 km^3 and represent one of the largest subaerial flood basalt ever to erupt on Earth. A very large volume may also remain in the uppermost crust in the form of dykes and sills. In addition, basalts of the ECMIP of North America, which most likely cause the East Coast Magnetic Anomaly (Kelemen and Holbrook, 1995), have a submarine area of about 60,000 km^2, with perhaps 1,300,000 km^3 of extrusive lavas. However, these basalts have not been yet been genetically connected to the continental CAMP and it remains a possibility that their formation was a different event, possibly younger, and possibly associated with the onset of seafloor spreading (Benson, 2003).

Whole-rock analyses of dykes, sills, and lavas of the CAMP tend to fall into three chemical groups, as outlined in McHone (2000) and used by Salters et al. (2003). These groups are characterised based on average values of TiO2: 0.62 % (low, or LTi), 1.26 % (intermediate, or ITi), and 3.21 % (high, or HTi), and other components such as magnesium, nickel, and various element ratios. All are tholeiites, with the LTi group mostly olivine normative, and ITi and HTi groups mostly quartz normative. As expected, phenocrysts of olivine tend to be abundant in the LTi dykes and sills, while minor interstitial quartz can be found in many of the ITi and HTi dykes, as well as early olivine in the larger intrusions.

There are also distinctions with respect to dyke swarm locations and orientations (Fig. 2). Dykes and sills of LTi basalt are nearly all found in basins and NW-trending dyke swarms in the SE USA, whereas most of the HTi dykes are on the margins of South America and Africa that were adjacent before rifting. They also tend to be in NW-SE trending dykes. LTi and HTi magmas are apparently not represented among the remnants of surface flows within the CAMP. The ITi dykes and sills are joined by large basalt flows preserved in rift basins of eastern North America and NW Africa. In those basin areas, the ITi dykes tend to trend NE-SW, but this group is very widespread and also has N-S dykes and other trends in other areas around the CAMP (Fig. 2).

Several localities in the SE USA show ITi dykes crosscutting LTi sills and dykes (Ragland et al., 1983) that are overall temporally overlapping/coeval (~201 Ma) with only minor variations (<0.5 Ma) (Hames et al., 2000; Blackburn et al., 2013). High-precision dates suggest about 570,000 years between the earliest and latest basin basalts (Olsen et al,. 2003), based on basin stratigraphy correlated with Milankovitch climatic cycles. The Triassic-Jurassic boundary occurs above the oldest ITi basalts in eastern North America (Cirilli et al., 2009), but the end-Triassic extinction horizon is still defined a meter or so beneath the oldest basin basalt (Olsen et al., 2003). Older basalts and large sills (Davies et al., 2017) exist in Morocco (Deenen et al., 2010) that precede the end-Triassic mass extinction for which it is now generally recognised that the CAMP is the prime causal candidate (Blackburn et al., 2013). The petrological diversity of CAMP basalts thus suggests considerable mantle-source heterogeneity and lithospheric influence on the magmas (Section 3.5).

3.3 Timing of Rifting and Magmatism

Although CAMP magmatism occurred in extremely intense but relatively brief episodes around 201 Ma, the tectonic activity that led to the breakup of Pangaea was much more prolonged (Frizon De Lamotte et al., 2015; Keppie, 2016). The oldest rift basin sediments around the central Atlantic are early Carnian (Late Triassic), possibly older than 230 Ma (Olsen, 1997). In the SE USA, rifting ended before CAMP magmatism, such that sediments and basalts are spread across wide areas rather than being controlled by subsiding basins (Schlische et al., 2003). Seismic reflection data suggests that younger Cretaceous strata are deposited directly upon the CAMP lava plains (McBride et al., 1989). In the NE rift basins, thick Early Jurassic sediments overlie basalts (Olsen, 1997), showing that rifting continued for 5 to 10 Myrs or more after the youngest CAMP flows, ceasing by the early Middle Jurassic (Schlische et al., 2003). This diachronous rifting was once thought to correspond to the changes in dyke orientations from south to north in eastern North America, but it is now known that the dyke magmas were roughly coeval.

The actual age of continental separation and production of the new ocean is uncertain and needs further research. It is generally assumed, and supported by seismic interpretation (Kelemen and Holbrook, 1995), that eruption of the thick seaward-dipping volcanic wedge along the eastern continental margin of North America immediately preceded the formation of Atlantic Ocean crust. However, the oldest drift sediments along the western Atlantic margin appear to be 179 to 190 Ma (Benson, 2003), or about the age of the youngest post-CAMP rift basin strata. There may thus be a 10-Myr gap between cessation of CAMP magmatism and seaward-dipping wedge magmatism and formation of new ocean crust.

3.4 Kinematics of the Central Atlantic rift – implications for breakup

Early Mesozoic dykes in eastern North America and NW Africa have been proposed to radiate from a central area at the Blake Plateau, near the modern-day Bahamas (May, 1971). This led to a model in which a deeply rooted thermal anomaly produced not only the dykes and basalts (Morgan, 1983) but also caused the rifting of Pangaea and the opening of the Central Atlantic Ocean (Storey et al., 2001). This model has been challenged by numerous previous workers (e.g., McHone, 2000).

McHone (2000) argued that the circum-Atlantic dykes are oriented parallel to segments of adjacent central Atlantic rifted margins (Fig. 2), and are not radial even within sets of regional dykes such as in the SE USA. Moreover, volcanic seamounts and islands of the Atlantic are much younger than breakup, so there is no volcanic plume track from the proposed centre evident, as would be required for such a mechanism (Pe-Piper et al., 1992; Pe-Piper et al., 2013). As described above, rifting that eventually opened the Central Atlantic started > 30 Myr before the magmatism, and the rift basins continued to develop for about another 10 Myr before tectonic activity shifted to the new ocean margins (Olsen, 1997). Thus, rifting was not contemporaneous with the massive production of CAMP basalts as expected for triggering by the arrival of a plume head.

Weigand and Ragland (1970) ascribed the chemical variations of the CAMP basalts to crystal fractionation within lithospheric magma chambers. However, it does not appear that all of the chemical variations observed in the CAMP magmas can be derived through differentiation or contamination of a common mantle melt (Salters et al., 2003). The upper mantle has substantial mineralogical, chemical, and temperature variations, or heterogeneous zones, which also influence composition (Shellnutt et al., 2017). The petrological diversity does not, however, support a model of a narrow mantle plume source (Tollo and Gottfried, 1989).

Components from crustal rocks that were subducted in much older plate collision events characterise most CAMP basalts (Merle et al., 2013; Puffer, 2001; Pegram, 1990). CAMP magmas are clearly derived from different compositions of sub-lithospheric mantle, some with substantial subduction contamination, in specific regions and across large geographic areas unrelated to any single centre. A preferred model for producing the CAMP is by the tectonic release of mantle melts that formed in a mantle warmed as a result of thermal insulation beneath the vast Pangaean supercontinent (Anderson, 1994; Merle et al. 2013). However, results of numerical models suggest that continental insulation is not the primary influence of supercontinents on mantle temperature (Heron and Lowman, 2010; 2014).

4.0 The breakup of East and West Gondwana

Breakup of East and West Gondwana during Early Jurassic times marked the end of the Gondwana supercontinent (Veevers, 2004; Klimke and Franke, 2016; Phethean et al., 2016)

(Fig. 3). Along the central region of the Gondwana rift, two oceanic basins record the tectono-magmatic history of the breakup. These are the West Somali Basin, from southern Somalia to northern Mozambique, and further south the Mozambique Basin, which is conjugate to the Riiser Larsen Sea/Lazarev Sea, Antarctica. Breakup followed a prolonged phase of episodic activity along the Karoo rift system and was closely contemporaneous with the eruption of the Karoo-Ferrar flood basalts and formation of the Lebombo volcanic monocline in Mozambique. Here, we discuss the spatio-temporal significance of tectonic and magmatic events, and their possible influence on breakup.

4.1 Overview East and West Gondwana rifting and breakup

Prior to the Middle Jurassic breakup of East and West Gondwana, tectonic activity along the Southern Trans-Africa Shear System, and much of the future line of continental separation in East Africa, had been underway since the Early Permian (Macgregor, 2018). Rifting associated with this early tectonism led to deposition of Karoo sediments along NW-SE and NE-SW trending basins during three main phases:

1) Extension between ~300 Ma and ~265 Ma along NW-SE trending basins and sinistral strike-slip along NE-SW trending basins resulted in sedimentation of rifts and local deposition within left-lateral step-over basins, respectively (e.g., Hankel, 1994).

2) A reconfiguration of the rift system occurred between ~259 and ~264 Ma with the onset of extension and rapid subsidence in NE-SW trending basins (Schandelmeier et al., 2004). Strike-slip deformation occurred along formerly extensional NW-SE trending basins (Delvaux, 2001).

3) Following the final episode, a brief pause in rifting occurred across most basins between ~249 to ~242 Ma (e.g., Hankel, 1994; Geiger et al., 2004; Frizon De Lamotte et al., 2015). This was followed by rejuvenation of rifting along NE-SW trending rifts (Schandelmeier et al., 2004), and little activity along NW-SE trending rifts (Delvaux, 2001). This rifting episode lasted until ~209 Ma (e.g., Hankel, 1994).

Deposition of Karoo supergroup sediments during these rifting phases was contemporaneous with development of the Cape Fold Belt in South Africa between 220 and 290 Ma (Frimmel et al., 2001). A link has been suggested between episodic development of the Karoo rift system (e.g., Hankel, 1994; Schandelmeier et al., 2004; Reeves et al., 2016) and compression across the Cape Fold Belt (Delvaux, 2001) which reactivated pre-existing basement weaknesses along the northern parts of the Karoo rift system (Reeves, 2014).

A long period of inactivity along the rift system then followed from ~209 Ma to ~183 Ma, after which many branches of the Karoo rift system along the line of future Gondwana separation reactivated in the Early Jurassic (~183 Ma) (Hankel, 1994; Papini and Benvenuti, 2008; Frizon De Lamotte et al., 2015). North of southern Tanzania, and south of northern Mozambique, Jurassic rifting overprints earlier Karoo rifts (Hunegnaw et al., 2007; Kassim et al., 2002; Catuneanu et al., 2005; Macgregor, 2018). The line of Jurassic continental breakup from southern Tanzania through northern Mozambique, however, shows little evidence of following an earlier Karoo rift system (e.g., Macgregor, 2018) and displays very different configurations

(Frizon De Lamotte et al., 2015). The distinct Jurassic rifting episode is clearly seen in southwestern Madagascar and southeast Tanzania (Geiger et al., 2004), where new half-grabens developed that crosscut Karoo rift structures and are filled by divergent wedges of Toarcian (Early Jurassic) syn-rift marine shales (Balduzzi et al., 1992; Macgregor, 2018).

4.2 Rifting and magmatism

The Jurassic rifting episode led to the final breakup of East and West Gondwana and was contemporaneous with major magmatism (Fig. 3). The Karoo LIP is primarily composed of the triple junction forming the Lebombo Monocline, the Okavango Dyke Swarm, and the Save-Limpopo Dyke Swarm centred on Mwenezi, Mozambique (e.g., Hastie et al., 2014). Other dyke swarms, sills, and significant flood basalts are preserved in Botswana and South Africa (Jourdan et al., 2005). The inner Explora Wedge and Ferrar LIP (ca. 183.6 ± 1 Ma; Encarnación et al., 1996) forms the Antarctic counterpart of the Karoo LIP.

The Lebombo Monocline may form part of the volcanic rifted margin of Mozambique and continental breakup is thought to have occurred along it (Klausen, 2009; Gaina et al., 2013). The monocline comprises progressively rotated dykes and seaward dipping lava flows, which are laterally segmented by scissor faults. This structure shows similarities to the North Atlantic volcanic rifted margins, and field relationships suggest that early tectonic extension became rapidly overwhelmed by dyke dialation (Klausen, 2009). As such, the Lebombo and Mwenezi volcanics may be the equivalent of SDR sequences (e.g., Davison and Steel, 2018), although the final location of continental breakup is still currently unresolved (e.g. Klausen, 2009).

To the east of the monocline, the Mozambique Plain is underlain by Mesozoic volcanics and basalts have been drilled in the Domo-1 well 300 km east of the Lebombo Monocline (e.g., Davison and Steel, 2018). However, it is uncertain if continental crust underlies these lavas. The final line of breakup may therefore have passed through Mwenezi, or failed here and instead passed around the Mozambique Plain. The Lebombo Monocline was formed over a long period of ~10 Ma (e.g., Jourdan et al., 2007; Hastie et al., 2014; Riley and Knight, 2001) from ~184 Ma to 174 Ma, with peak activity between 183-178 Ma (Hastie et al., 2014). This is ~3 Myrs earlier than the counterpart Ferrar magmatism on the conjugate Antarctica margin (Riley and Knight, 2001).

Lateral magma flow within the Lebombo Monocline and Okavango Dyke swarm is consistent with a magma source at the nearby Mwenezi triple junction (Hastie et al., 2014). However, the significant magmatism away from the Mwenezi triple junction, which additionally shows magma flow directions inconsistent with a Mwenezi origin, suggest additional sources of magmatism away from the triple junction (Hastie et al., 2014). The triple junction's NE branch, the 070° trending Save-Limpopo dyke swarm, was under orthogonal NNW-SSE extension during its intrusion (Le Gall et al., 2005). In addition, it has been demonstrated that the NW branch, the 110° trending Okavango Dyke Swarm, opened with transtensional dyke intrusion and was also under the same NNW-SSE stress field. Thus, the triple junction structure did not result from active extensional forces radiating from Mwenezi (Le Gall et al., 2005). The magmatism instead followed pre-existing lithospheric structures, in this case alongside an ESE-trending Proterozoic dyke swarm.

Approximately 10% of dykes in the Okavango swarm are Proterozoic, whilst the remaining 90% are Jurassic. Dykes of both ages show a strong geochemical affinity to each other, leading Jourdan et al. (2009) to suggest that both magmatic episodes were sourced from an enriched shallow mantle lithospheric source. Variations in magma composition in the Karoo LIP between low- and high-Ti magmas correlate with Proterozoic and Archean basement (Hawkesworth et al., 1999). Luttinen (2018) proposed an alternative bilateral division of magmas, into subduction and plume-related geochemical affinities, based on relative Nb abundance. There is no evidence for concurrent uplift during magma emplacement (Watkeys, 2002), and magmas young progressively from south to north (Jourdan et al., 2005), i.e. towards the Mwenezi triple junction.

Breakup-related volcanics at the continental margins of the Mozambique Basin, and its conjugates, the Lazarev Sea and the Riiser-Larsen Sea in Antarctica, comprise SDRs and high-velocity lower crustal bodies (Hinz et al., 2004; Leinweber and Jokat, 2012; Mahanjane, 2012; Mueller and Jokat, 2017). However, the volcanics terminate before the Mozambique Strait between Madagascar and Mozambique (Klimke et al., 2018). In the West Somali Basin to the north, there is little evidence for magmatism during the breakup and the basin is thought to be magma-poor (Coffin et al., 1986; Klimke and Franke, 2016; Phethean et al., 2016; Stanton et al., 2016; Stanca et al., 2016).

Despite the many plate kinematic models of breakup of Gondwana along the East African margin (Rabinowitz et al., 1983; Cox, 1992; Reeves et al., 2004; Eagles and König, 2008; Leinweber and Jokat, 2012; Gaina et al., 2013; Nguyen et al., 2016; Phethean et al., 2016; Davis et al., 2016; Reeves et al., 2016), the exact ages of formation of the West Somali and Mozambique basins are still poorly constrained. This is mainly because, if present, the earliest oceanic crust formed during the Jurassic Magnetic Quiet zone, where rapid polarity changes in the Earth's magnetic field resulted in seafloor spreading anomalies that are difficult to detect (Tominaga et al., 2008). The extinct spreading axis has been tentatively identified using gravity data from the West Somali Basin (Sauter et al., 2016; Davis et al., 2016; Phethean et al., 2016) but the identification of seafloor-spreading-related magnetic anomalies are still an active area of research.

In the West Somali Basin, Davis et al. (2016) identified magnetic anomalies as old as M24Bn (152.43 Ma). Gaina et al. (2013) suggest magnetic anomaly M40ny/M41 (~166 Ma) is the oldest and M2 (~127 Ma) is the youngest magnetic anomaly in the West Somali Basin, extending shorter periods suggested by Rabinowitz et al. (1983) (M10-M25; ~155 Ma to 134 Ma) and Segoufin and Patriat (1980) (M0-M21; ~147 Ma to 124 Ma). In addition, the stratigraphic record from the basin shows an overwhelming to marine sedimentation in the Early Bajocian at around 170 Ma (Coffin and Rabinowitz, 1992), in agreement with a breakup unconformity in the Morondava Basin at this time (Geiger et al., 2004).

Using new geophysical data Mueller and Jokat (2017) and Leinweber and Jokat (2012) tentatively identify M38n.2n or M41n (~164 or 165 Ma) as the oldest magnetic anomaly in the Mozambique Basin, extending earlier-identified seafloor-spreading anomalies M2 to M22 (~148-127 Ma; Segoufin, 1978; Simpson et al., 1979). However, in the conjugate Riiser-Larsen Sea, the oldest magnetic anomaly identified so far is M25n (~154 Ma) (Leitchenkov et al., 2008; Leinweber and Jokat, 2012). The Rooi Rand dyke swarm of the southern Lebombo Monocline has E-MORB geochemical affinity, has been dated at ~173 Ma (Jourdan et al., 2007a; Hastie et al., 2014). It is thought to reflect incipient breakup and early seafloor spreading in the southern Mozambique Basin. These records suggest that breakup in the Mozambique Basin occurred between ~173 Ma and 164.1 Ma, similar to the proposed age of breakup in the West Somali Basin of ~170-152 Ma (references in previous paragraph).

4.3 Timing of rifting and magmatism

The Karoo LIP erupted in Botswana and South Africa from 185 Ma to 178 Ma (Jourdan et al., 2005). Magmatic ages within the Lebombo Monocline, and the Okavango and Save-Limpopo Dyke Swarms overlap each other significantly and lie in the range 183 Ma to 174 Ma (Hastie et al., 2014). If the onshore Lebombo Monocline is in fact a volcanic rifted margin, this would indicate a significant overlap in flood basalt generation and incipient lithospheric breakup of the Mozambique Basin. In view of the south-to-north age progression of the Karoo flood basalts and sills in Botswana and South Africa it is appropriate to compare magmatic ages from the flood basalts and volcanic margins from similar latitudes. Both the Northern Flood Basalt Province described by Jourdan et al. (2005) and the Northern Lebombo Dyke Swarm lie at approximately the same latitude, and were intruded between 182 Ma and 178 Ma (Hastie et al., 2014). Advanced lithospheric extension along the volcanic rifted margin near the Northern Lebombo Dyke Swarm may therefore have already been already present at the time of latitudinally equivalent flood basalt eruption. If, however, the Lebombo Monocline is not the volcanic rifted continental margin, then it is apparent that the LIP volcanism has no spatial relationship with continental breakup that occurred about 200 km farther east.

4.4 Kinematics of the East and West Gondwana breakup – implications for breakup

While earlier studies proposed that the Mozambique Basin and West Somali Basin opened in a generally N-S direction, more recent plate tectonic reconstructions argue for an almost simultaneous opening of both basins in a NW-SE direction (e.g., Gaina et al., 2013; Klimke and Franke, 2016; Phethean et al., 2016; Reeves et al., 2016). This is supported by the stress configuration derived from dyke swarms of the Karoo LIP emplaced during the Jurassic rift phase (Le Gall et al., 2005). Several dyke swarms record a NNW-SSE initial opening direction during the Jurassic (Le Gall et al., 2005).

NNW-SSE gravity lineaments related to spreading in proximity to the African coast have been identified within the Western Somali Basin, between Madagascar and Africa (Davis et al., 2016; Phethean et al., 2016). This newly identified phase of NNW-SSE spreading lasted between ~170 Ma and ~153 Ma and is consistent with the initial NNW-SSE opening of the Mozambique Basin (Phethean et al., 2016). NNW-SSE spreading was superseded by N-S spreading from ~153 Ma following the passing of Madagascar beyond the continental lithosphere of Mozambique and development of the Davie Fracture Zone (Reeves, 2017) along which East Gondwana was transposed.

Reeves et al. (2016) performed plate tectonic reconstruction and found initial NW-SE motion is required. The pole of rotation between East and West Gondwana during the early phase of separation (~183 Ma to ~153 Ma) lay ~2000 km west of the SW tip of present-day Africa. This pole location requires that extension rates across the Gondwana rift increase to the NE. As the time of breakup is primarily a function of cumulative extension across a rift, this would result in SW-propagating breakup between East and West Gondwana. Such a structural configuration is generally supported by the sedimentological record (Salman and Abdula, 1995) and would indicate that the rift propagated towards the Mwenezi triple junction.

Understanding of the timing and kinematics of the Western Somali Basin, and to the south of this the Mozambique Basin and its conjugate the Riiser Larsen Sea/Lazarev Sea, Antarctica, is still incomplete. As a result it is difficult to derive a final and conclusive model about the relationship between the Jurassic breakup of East and West Gondwana, including the formation of the Mwenezi triple junction and the Karroo-Ferrar LIP. However, it is clear that the triple junction structure was not the result of active extensional forces radiating from Mwenezi, and magmatism instead followed pre-existing lithospheric structures. The massive magmatic extrusion that formed the Karoo-Ferrar LIP likely predates rifting but breakup and formation of the oceanic basins did not initiate close to the triple junction. It is therefore more likely that the rift and subsequent breakup migrated towards the triple junction. Nevertheless, work is still needed to fully understand the relationship between the Jurassic breakup of East and West Gondwana and the formation of the Mwenezi triple junction and the Karroo-Ferrar LIP.

5.0 The opening of the South Atlantic

In the Early Cretaceous, West Gondwana, a southern constituent of Pangaea, broke up to form South America and Africa with continuous spreading resulting in the sustained expansion of the South Atlantic Ocean (Rabinowitz and Labrecque, 1979; Ben-Avraham et al. (1997) Lawver et al., 1998; Jokat et al., 2003; Eagles 2007; Moulin et al., 2009; Lovecchio et al., 2018) (Fig. 4). The contemporaneous Paraná–Etendeka continental flood-basalt provinces in Brazil and Namibia, respectively, are frequently attributed to an Early Cretaceous Tristan da Cunha plume with the Walvis Ridge and Rio Grande Rise comprising plume tail magmatism (Morgan, 1981; Peate, 1997). As discussed herein, there are significant spatial and temporal mismatches between the proposed plume and these structures.

5.1 Overview of South Atlantic rifting and breakup

Regardless of the remarkable geometrical fit between the rifted continental margins of South America and Africa (Fig. 4), systematically initially investigated by Wegener (1915) and by numerous workers since (e.g., Gladczenko et al., 1997; Granot and Dyment, 2015), both the rift and breakup phases were complex, with evidence of multiple stages of rifting (Lovecchio et al., 2018), and the possible influence of structural inheritance (Ben-Avraham et al., 1997; Salomon et al., 2015).

Continental extension may have begun in isolated centres in South America during the Late Triassic (at about 210 Ma) when almost all parts of south and west Gondwana were affected by magmatism resulting in high heat flow (Macdonald et al., 2003). In addition to this Late Triassic to Early Jurassic rifting phase, there was a Middle Jurassic extensional phase lasting almost 40 Ma, from Valanginian to late Albian time (Early Cretaceous), that completed separation of Africa and South America to separate completely (Keeley and Light, 1993; Szatmari, 2000). The line of continental separation and the position of the principal failed rifts were controlled by the position of boundaries between different aged basement and the inheritance of basement structural grain (Macdonald et al., 2003). Breakup is reasonably well understood but location and magnitude of continental intraplate deformation during rifting, particularly affecting South America, requires further work (see e.g. Eagles, 2007; Heine et al., 2013; Moulin et al., 2009; Torsvik et al., 2009).

5.2 Rifting and magmatism

Continental breakup and initial seafloor spreading in the South Atlantic were accompanied by extensive transient magmatism as inferred from sill intrusions, flood basalt sequences, and voluminous volcanic wedges and high-velocity lower crust at the present continental margins. Voluminous volcanism affected both Mesozoic intracratonic basins onshore (Paraná-Etendeka flood-basalt province; Peate, 1997; Renne et al., 1992; Trumbull et al., 2007; Foulger, 2017) and the rifted crust offshore (Bauer et al., 2000; Franke et al., 2007; Gladczenko et al., 1997; Gladczenko et al., 1998; Hinz et al., 1999; Koopmann et al., 2014; Mohriak et al., 2008; Paton et al., 2016; Stica et al., 2014) (Fig. 4).

Menzies et al. (2002) and Moulin et al. (2009) compiled published geochemical data and radiometric dates for the dykes and the lava flows of the Paraná–Etendeka flood-basalt provinces. According to these compilations, volcanic activity peaked in the late Hauterivian – early Barremian (Early Cretaceous; 133-129 Ma, and 134–130 Ma, respectively). Apart from the age of the basalts, there is controversy about the source of Paraná–Etendeka magmas (see e.g. Renne et al., 1992; Peate, 1997; Hawkesworth et al., 1999; Trumbull et al., 2007; Rocha-Júnior et al., 2013; Comin-Chiaramonti et al., 2011; Will et al., 2016; Foulger, 2017).

The Early Cretaceous opening of the southern South Atlantic took place between 135 to 126 Ma (Heine et al., 2013; Moulin et al., 2009; Macdonald et al., 2003; Rabinowitz and Labrecque, 1979). Multichannel seismic and potential field data suggest the oldest magnetic chron in the southern South Atlantic related to oceanic spreading is M9 (ca. 135 Ma) Moulin et al., 2009). Older anomalies, previously identified as M11 (ca. 137 Ma), are found within the SDRs (Koopmann et al., 2016; Corner et al., 2002). There is still some uncertainty about the age of the first oceanic crust near the Falkland Plateau, where strike-slip deformation from the Falklands-Agulhas fracture zone hampers identification of the earliest spreading anomalies. Collier et al. (2017) and Hall et al. (2018) identified M10r (134.2 Ma, late Valanginian) as the oldest recognisable chron at the southern tip of the South Atlantic. This agrees with the suggestion of Becker et al. (2012) that the breakup unconformity, identified in rift basins at the northern edge of the Falkland plateau, is contemporaneous with the well-dated rift-to-sag unconformity in the North Falkland Basin. This indicates a Valanginian (~135 Ma; Early Cretaceous) age for the first oceanic crust in the southern South Atlantic.

Most of the southern South Atlantic continental margins are volcanic (Gladczenko et al., 1997; Becker et al., 2014; Foulger, 2017) (Fig. 4). However, the southernmost 400-km-long portion lacks SDRs (Koopmann et al., 2014b; Becker et al., 2012; Franke et al., 2010; Hall et al., 2018). Thus, from the magnetic anomalies seaward of the SDRs, volcanic rifting onset abruptly, shortly before 137 Ma (Koopmann et al., 2016). From there towards the north, the progressive continental breakup was accompanied by large-scale transient magmatism with the formation of voluminous SDR wedges and high-velocity lower crustal bodies over the ~1800 km to the Florianopolis/Rio Grande fracture zones offshore Namibia/Brazil (Becker et al., 2014). The SDRs were emplaced consecutively northward, as indicated by the progressive termination of the pre-M4 magnetic seafloor spreading anomalies within the volcanic wedges. Only from magnetic chron M4 (ca. 130 Ma) onward was oceanic crust formed across the entire southern South Atlantic (Koopmann et al., 2016).

Although magnetic anomalies from M4 (~130 Ma) onwards have been proposed for the central South Atlantic, north of the Florianopolis (or Rio Grande) fracture zone (Bird and Hall, 2016), most authors agree that breakup was delayed (by 10-20 Myr) across this fracture zone (Torsvik et al., 2009; Moulin et al., 2009; Quirk et al., 2013; Heine et al., 2013). At the latitude of the Paraná–Etendeka flood-basalt provinces, rift propagation was apparently blocked. At this position, one of the fundamental structures in the South Atlantic development (Moulin et al., 2013), the Florianópolis (or Rio Grande) fracture zone, is found. This fracture zone hosted significant offset during breakup (150 km; Elliott et al., 2009). To its north, the central South Atlantic is characterised by minor SDRs which were deposited contemporaneously with Aptian salt deposits (Mohriak et al., 2008). A number of aborted rifts developed along the Brazilian margin (the Campos, Santos, and Esperito Santos Basins) and the crust was extremely stretched and thinned before the two spreading axes in the central and southern South Atlantic connected (Mohriak et al., 2002; Evain et al., 2015).

Sporadic but widespread magmatic activity continued well after breakup (80 Ma and younger) in southern Africa and Brazil (Comin-Chiaramonti et al., 2011). This magmatism is most commonly manifest as alkaline intrusions, which are locally numerous (e.g., kimberlite fields) but smaller in volume than the Early Cretaceous activity.

5.3 Timing of rifting and magmatism

A key question is the relative timing of extension and emplacement of the large-volume magmatic flows, both onshore (Paraná–Etendeka flood-basalts) and offshore (SDRs). The best estimate currently available for the onset of rifting adjacent to the Walvis Ridge/Rio Grande Rise is about 134-135 Ma (Bradley, 2008; Moulin et al., 2009). This preceded surface breakup in the immediate vicinity. In both provinces, the basalts were deposited in north-south–trending rift basins, showing that rifting preceded flood volcanism, however (Clemson et al., 1997; Glen et al., 1997). The Paraná–Etendeka flood-basalts also erupted at the intersection of a major, activated, transverse extensional structure with the developing line of breakup (Foulger, 2017). Numerical modelling suggests that depth-dependent extension was underway for a considerable period before surface rupture. This is in line with the magma flow directions of both the basaltic rocks from the Etendeka igneous province of Namibia and from the Paraná province in Brazil.

Magnetic seafloor spreading anomalies indicate that the peak magmatism (~132 Ma) of the Paraná–Etendeka flood-basalts postdates emplacement of SDRs in the southern South Atlantic (Koopmann et al., 2016). Only if the M-sequence geomagnetic polarity timescale is used (Malinverno et al., 2012), instead of the popular Gradstein and Ogg (2012) timescale, does dating suggest that the SDRs were emplaced simultaneously (Koopmann et al., 2016). As the

SDRs mark the final stage of continental rifting it is evident that the complete extensional phase and likely also earliest seafloor spreading in the southern South Atlantic predate the emplacement of the Paraná and Etendeka basalts (Franke, 2013).

5.4 Kinematics of the South Atlantic rift – implications for breakup

The South Atlantic opened by south-to-north propagation (Gaina et al., 2013; Heine et al., 2013; Seton et al., 2012; Moulin et al., 2009; Jokat, 2003; Macdonald et al., 2003; Austin and Uchupi, 1982; Rabinowitz and Labrecque, 1979) (Fig. 4). As pointed out by Franke (2013), this opening direction contradicts the hypothesis that rifting migrated away from the Paraná–Etendeka flood-basalt provinces (Fig. 4). On the contrary, rifting migrated towards it, at odds with a model whereby continental breakup was triggered by an active upwelling mantle plume currently beneath the Tristan da Cunha hotspot. Other candidate mechanisms must therefore be sought as a trigger for breakup.

When reconstructing the South Atlantic, the Cape fold belt in South Africa aligns with the Ventana (or Sierras Australes) Hills in Argentina. Paton et al. (2016) identify the South African Cape fold belt offshore South Africa and propose that initial rifting along western Gondwana was a consequence of extensional reactivation of the western Gondwanan Fold Belt (Fig. 4). The rift basins are thought to have formed through gravitational collapse of the fold belts such that rift basin geometry was controlled by underlying fold belt geometry. This resulted in broadly SW-orientated (with respect to Africa) extension in Argentina/South Africa. According to Paton et al. (2016), during the mid-Cretaceous, the rift configuration changed significantly and extension followed a north–south trend, i.e. perpendicular to the fold-belt. This geometry fits well with the proposed earlier clockwise rotation of extensional deformation throughout the Early Cretaceous based on structural data from the continental margins(Franke, 2013).

The highly asymmetric subequatorial margins of Brazil and West Africa almost certainly did not rift apart in a pure-shear fashion and simple-shear rifting mechanisms have been suggested (Mohriak et al., 2008). In addition, it has been suggested that the structure and shape of the continental margins show considerable deviations from symmetric structures expected from active rifting, triggered by a plume below the rift (Geoffroy, 2005; Campbell and Kerr, 2007). However, if there was a plume, the style and shape of breakup would still be governed or at least influenced by inherited lithospheric structures, so the margins could still have any kind of complexities, including asymmetry. With respect to volcanics, high-velocity lower crust, dyke orientations, and fault patterns, the complementary southern South Atlantic rifted margins experienced distinct asymmetric evolution during breakup (Salomon et al., 2017; Koopmann et al., 2016; Becker et al., 2016; Becker et al., 2014). The asymmetry in offshore magmatism with considerably more SDRs and volume of high-velocity lower crust on the African margin is surprising, given the opposite asymmetry in the onshore Paraná–Etendeka flood-basalt provinces. On the basis of fission-track and denudation studies on both margins, an explanation in greater post-rift uplift and erosion on the African margin has been ruled out (Becker et al., 2014). Instead, South America offered more favourable structures for magma ascent and extrusion than South Africa. This supports mainly passive rifting as proposed earlier by Maslanyj et al. (1992).

A seismic refraction study at the easternmost Walvis Ridge, including the junction with the Namibian coast, found a small intruded area around the Walvis Ridge (Fromm et al., 2015). Also onshore, in the landfall area of the Walvis Ridge at the Namibian coast, a narrow region (<100 km) of high-seismic-velocity anomalies in the middle and lower crust, interpreted as a massive mafic intrusion, has been identified by seismic reflection and refraction data (Ryberg et al., 2015). These data and observations are not particularly consistent with a broad thermal plume head beneath the opening South Atlantic.

To the north of Walvis Ridge, the abrupt disappearance of SDRs (Elliott et al., 2009) accompanies a dramatic decrease in crustal thickness from 35 km below Walvis Ridge to 5–6 km crust in the central South Atlantic (Fromm et al., 2015). A similar sudden disappearance of SDRs occurs south of a major transfer zone in the southern South Atlantic (Koopmann et al., 2014b; Becker et al., 2012). These abrupt changes in magmatic volume are also inconsistent with a large-scale thermal source in the sublithospheric mantle as an origin for the magmatism. Gradual variations of mantle properties and dynamics are expected to generate smooth transitions over at least a hundred or a few hundreds of kilometres, not sharp transitions (Franke et al., 2010).

In addition, the architecture of the SDRs implies an episodic emplacement with multiple magmatic phases alternating with magma-starved phases (Franke et al., 2010). The South Atlantic unzipped in jumps from south to north and the SDRs were emplaced consecutively along the successive northward propagating rift zones (Clemson et al., 1997; Franke et al., 2007; Koopmann et al., 2014; Stica et al., 2014). Between the Falkland-Agulhas fracture zone and the Walvis Ridge/Rio Grande Rise (Fig. 4), this process lasted for approximately 5 Myrs as shown by the earliest magnetic chrons in the South Atlantic (Koopmann et al., 2016; Hall et al., 2018).

6.0 Opening of the NE Atlantic, the Labrador Sea and Baffin Bay

The northern North Atlantic realm contains two main spreading branches (Vogt and Avery, 1974) (Fig. 5). The Labrador Sea – Baffin Bay system (here referred to as the NW Atlantic as in Abdelmalak et al., 2018) separated Greenland and North America (Vogt and Avery, 1974; Srivastava, 1978; Torsvik et al., 2002; Hosseinpour et al., 2013; Peace et al., 2016; Welford et al., 2018). Subsequently, the NE Atlantic began to open, separating Greenland and Europe (Talwani and Eldholm, 1977; Skogseid et al., 2000; Lundin and Doré, 2005b; Le Breton et al., 2012; Gaina et al., 2009; Gernigon et al., 2015; Gaina et al., 2017a; Gaina et al., 2017b; Schiffer et al., 2018; Foulger et al. this volume). A complex junction exists between these branches to the north of the Charlie-Gibbs Fracture Zone (CGFZ) (Gaina et al., 2009) (Fig. 5). Switchover from the western spreading ridge to the eastern ridge was one of the most significant events in the evolution of the North Atlantic (Nielsen et al., 2007; Jones et al., 2017). Understanding the mechanisms that drove this switchover remains one of the most important unresolved questions in understanding North Atlantic tectonics (Peace et al., 2017a).

In addition to these first-order spreading axes, Northeast Atlantic oceanic crust is further structurally divided in proximity to Iceland by the Kolbeinsey and Aegir ridges (Fig. 5). The genesis of Iceland, and the proximal Jan Mayen Microplate Complex (JMMC) still present unresolved questions (e.g., Müller et al., 2001; Foulger and Anderson, 2005; Gernigon et al., 2015; Blischke et al., 2016; Schiffer et al., 2018b; Blischke et al., 2019; Schiffer et al. this volume). Regions where major extension occurred without breakup (i.e. failed rifts and transforms), must be accounted for in geodynamic models. These include the Davis Strait (Suckro et al., 2013; Peace et al., 2018b), the North Sea (Cowie et al., 2005), the Rockall Basin (Shannon et al., 1994; Roberts et al., 2018), and the Hatton Basin (Hitchen, 2004) and potentially also the Greenland-Iceland-Faeroes Ridge (GIFR) (Foulger et al. this volume). In addition, diffuse intracontinental deformation may also have been associated with breakup (e.g., the Eurekan Orogeny; Nielsen et al., 2007; Nielsen et al., 2014; Heron et al., 2015; Piepjohn et al., 2016; Schiffer and Stephenson, 2017; Gion et al., 2017; Stephenson et al. this volume). Although difficult to quantify, these events must be accounted for in models of the breakup of the North Atlantic (Ady and Whittaker, 2018).

6.1 Overview of North Atlantic rifting and breakup

Prior to breakup, the proto-North Atlantic region comprised an assemblage of Archaean and Proterozoic terranes (Kerr et al., 1996; St-Onge et al., 2009; Štolfová and Shannon, 2009; Engström and Klint, 2014; Grocott and McCaffrey, 2017; Schiffer et al. this volume). Understanding the pre-breakup extensional phases and orogenies is crucial to understanding Mesozoic-Cenozoic breakup because of the clear influence of structural inheritance (Dore et al., 1997; Schiffer et al., 2015; Peace et al., 2018a; Peace et al., 2018b; Schiffer et al., 2018a; Phillips et al., 2018; Rotevatn et al., 2018; Gernigon et al., 2018; Schiffer et al. this volume).

Following the collision of Laurentia, Baltica and Avalonia in the Ordovician and Silurian (Roberts, 2003; Gee et al., 2008; Leslie et al., 2008), and subsequent gravitational extensional collapse (Dewey, 1988; Dunlap and Fossen, 1998; Rey et al., 2001; Fossen, 2010), the North Atlantic region may have experienced phases of lithospheric delamination and associated uplift for 30–40 Ma followed by a long period of rifting (Andersen et al., 1991; Dewey et al., 1993). The North Atlantic margins, including the Labrador Sea and Baffin Bay, experienced multiple phases of extension between the Devonian collapse of the Caledonian Orogen (Roberts, 2003) and early Cenozoic break-up (Srivastava, 1978; Doré et al., 1999; Lundin and Doré, 2018).

Multiple pre-breakup rift phases are documented in the stratigraphic record of both the NE and NW Atlantic (Umpleby, 1979; McWhae et al., 1980; Srivastava, 1978; Lundin, 2002; Oakey and Chalmers, 2012; Barnett-Moore et al., 2016; Nirrengarten et al., 2018). Rifting started as early as the Permian, was widespread during the Triassic, and continued into the Jurassic Cretaceous, and Cenozoic (Umpleby, 1979; Stoker et al., 2016). In the NE Atlantic, an early rifting pulse from Late Permian to earliest Triassic is expressed regionally in the stratigraphic record. These Permian–Triassic successions record a northward transition from an arid interior setting to a passively subsiding mixed-carbonate siliciclastic shelf margin (Stoker et al., 2016). In the Early Jurassic, the sedimentary record shows thermal subsidence and mild extensional tectonism (Stoker et al., 2016). In the Late Jurassic, the stratigraphic record reveals an intense phase of rifting across most of the NE Atlantic. Cretaceous sections record predominantly marine strata deposition within broad zones of extension (Stoker et al., 2016).

Following prolonged regional rifting (Larsen et al., 2009; Stoker et al., 2016), propagation of the Central Atlantic into the proto-North Atlantic began in Early Aptian time (e.g., Lundin, 2002; Barnett-Moore et al., 2016). The propagating spreading centre produced the oldest oceanic crust of the North Atlantic and is marked by the M0 magnetic anomaly (121-125 Ma; Malinverno et al., 2012) offshore Iberia and Newfoundland (Lundin, 2002; Tucholke et al., 2007; Eddy et al., 2017). By the Late Aptian (Early Cretaceous), spreading reached the Galicia Bank (Boillot and Malod, 1988). This was followed by formation of the Bay of Biscay triple junction in the Late Aptian or Early Albian (Early Cretaceous) where spreading continued until the Late Cretaceous (Williams, 1975). From the latest Cretaceous to the Eocene, however, the NW movement of Iberia with respect to Eurasia caused the Bay of Biscay to partly subduct beneath Iberia, forming the Pyrenees (Boillot and Malod, 1988). From the Bay of Biscay triple junction spreading propagated NW and reached the Goban Spur in Middle to Late Albian time (e.g., Tate, 1993). By the Santonian (Late Cretaceous), breakup had reached the Charlie Gibbs Fracture Zone (CGFZ) and significant extension, occurred in the Rockall Basin during the Cretaceous (Shannon et al., 1994; Hitchen, 2004).

The NW Atlantic was the next region to open (Srivastava, 1978; Chalmers and Pulvertaft, 2001; Lundin, 2002; Hosseinpour et al., 2013; Keen et al., 2017; Oakey and Chalmers, 2012; Abdelmalak et al., 2018; Welford et al., 2018). This extinct spreading system comprises the Labrador Sea in the south and Baffin Bay in the north (Fig. 5) (Chalmers and Pulvertaft, 2001). These are connected via the Ungava Fault Zone, a transform fault system running through the Davis Strait bathymetric high (Suckro et al., 2013; Peace et al., 2017a; Peace et al., 2018c). The Labrador Sea, Davis Strait and Baffin Bay formed via multiphase, divergent motion between Greenland and North America (e.g., Chalmers and Pulvertaft, 2001; Hosseinpour et al., 2013). Rifting prior to breakup occurred from at least the Early Cretaceous, but potentially as early as the Triassic according to dykes in southwest Greenland (Larsen et al., 2009; Secher et al., 2009) and, with some uncertainty, Labrador (Wilton et al., 2002; Tappe et al., 2006; Tappe et al., 2007; Wilton et al., 2016; Peace et al., 2016).

Onset of spreading in the Labrador Sea is thought to have occurred in the Early Campanian (Chron 33; ca. 80 Ma) (Roest and Srivastava, 1989; Srivastava and Roest 1999). In contrast, Chalmers and Laursen (1995) propose that Chrons 33 and 27 represent transitional crust with true oceanic crust in the Labrador Sea first generated in the Palaeocene (Chron 27; ca. 62 Ma). Keen et al. (2017), however, state that the ocean-continent boundary lies near magnetic anomaly Chron 31 (ca. 68 Ma), and divide the oceanic region into inner and outer domains, which merge near magnetic Chron 27 (ca. 62 Ma). The outer domain of Keen et al. (2017) is interpreted as steady-state seafloor spreading with well-developed linear magnetic anomalies, while the igneous crust of the older, inner domain is generally thinner, and more variable.

During the separation of Greenland and North America, oceanic crust was not formed in the Davis Strait (Suckro et al., 2013; Peace et al., 2017b), in part because of the primarily strike-slip nature (Wilson et al., 2006; Peace et al., 2018c). In Baffin Bay, oceanic spreading probably also occurred simultaneously with spreading in the Labrador Sea. This is, however, uncertain and oceanic crust there is undoubtedly more limited (Jackson et al., 1979; Hosseinpour et al., 2013). Regardless of the existence of older oceanic crust in the Labrador

Sea, it is generally accepted that Early Eocene (Chron 24; ca. 54 Ma) oceanic crust floors Baffin Bay (e.g., Chalmers and Pulvertaft, 2001).

Events in the NW Atlantic may be linked to changes in plate kinematics in the NE branch of the Atlantic (Gaina et al., 2009). During the Early Eocene (Chron 24; ca. 54 Ma), seafloor spreading began in the NE Atlantic, marking a major tectonic reorganisation (Lundin, 2002; Nielsen et al., 2007; Mosar et al., 2002; Gaina et al., 2016). The direction of spreading in the Labrador Sea and Baffin Bay system rotated to NNE-SSW (e.g., Abdelmalak et al., 2012; Peace et al., 2018a). This slowed seafloor spreading that was oblique to the earlier ridge system (Hosseinpour et al., 2013). A triple junction formed between the Labrador Sea, the NE Atlantic, and the southern North Atlantic, which was active until spreading ceased in the Labrador Sea in the earliest Oligocene (Chron 13; ca. 35 Ma) (e.g., Srivastava & Roest 1999). In the NE Atlantic, the abnormal thickness of the oceanic crust initially produced (ca. 54 Ma) decreased and a steady state was reached by the Middle Eocene (ca. 48 Ma) (Holbrook et al., 2001b; Lundin and Doré, 2005b; Storey et al., 2007; Mjelde and Faleide, 2009). By ca. 36-32 Ma, spreading had entirely relocated to the NE Atlantic (Roest and Srivastava, 1989; Barnett-Moore et al., 2016) and terminated along the Labrador Sea-Baffin Bay axis (Chalmers and Pulvertaft, 2001). Greenland then became part of the North American plate (Oakey and Chalmers, 2012; Barnett-Moore et al., 2018).

In the Norwegian Sea of the NE Atlantic, development of sea-floor spreading along the Reykjanes, Mohns, Ægir and Kolbeinsey ridges is relatively well understood (e.g., Lundin, 2002; Gernigon et al., 2015; Blischke et al., 2016; Zastrozhnov et al., 2018). The Ægir Ridge may represent the southern tip of a southward-propagating Arctic rift system that migrated west to form the Kolbeinsey Ridge. This was a transitional process with delocalisation starting at ~40 Ma and the Ægir Ridge becoming extinct sometime between ca. 21 and 28 Ma (e.g., Lundin, 2002). The overlapping geometry of the Ægir and Kolbeinsey Ridges was maintained during the subsequent sea-floor spreading (Müller et al., 2001; Schiffer et al., 2018). The Kolbeinsey Ridge linked with the Mohns Ridge, via the West Jan Mayen Fracture Zone in earliest Oligocene time (Chron 13; ~33 Ma) as indirectly dated by the eastern termination of the West Jan Mayen Fracture Zone, which reaches Chron 13 on the east side of the Mohns Ridge. This link made further spreading along the Ægir Ridge redundant and seafloor spreading ceased along the Ægir Ridge at approximately Chron 12 (Jung and Vogt, 1997). The link between the Kolbeinsey and Mohns ridges represents the linkage between the Arctic and Atlantic oceans (Lundin, 2002).

6.2 Rifting and Magmatism

Rifting and breakup of the northern North Atlantic was accompanied by significant, widespread magmatism (Eldholm and Grue, 1994; Mjelde et al., 2008; Hansen et al., 2009; Nelson et al., 2015; Wilkinson et al., 2016; Á Horni et al., 2017; Clarke and Beutel, this volume) (Fig. 5). This was particularly abundant during and after breakup (Saunders et al., 1997; Hansen et al., 2009; Wilkinson et al., 2016; Á Horni et al., 2017), although some magmatism also occurred during the preceding rifting (e.g., Larsen et al., 2009; Wilkinson et al., 2016). The continental passive margins of the southern North Atlantic (e.g. Newfoundland – Iberia and Labrador – southwest Greenland) are typically considered to be magma-poor (Chalmers, 1997; Chonian et al., 1995; Chalmers and Pulvertaft, 2001; Whitmarsh et al., 2001; Keen et al., 2017), whereas the margins further north (e.g., East Greenland, the NW European margin, and Central West Greenland) are considered to be 'magma-rich', and to contain SDRs and HVLCBs (Geoffroy et al., 2001; Breivik et al., 2012; Keen et al., 2012; Magee et al., 2016; Petersen and Schiffer, 2016; Larsen et al., 2016).

An early, coherent magmatic province in the North Atlantic realm was the Permo-Carboniferous Skagerrak LIP found in southern Sweden and Norway, Denmark, northern-central Europe and the British Isles (Heeremans et al., 2004; McCann et al., 2006). This igneous province was coeval with a general period of tectonic unrest and magmatic hyperactivity in Europe, possibly connected to the collapse of the Variscides that might have included extreme lithospheric thinning and delamination (Doblas et al., 1998; Timmerman et al., 2009; McCann et al., 2006; Meier et al., 2016).

Pre-breakup magmatism, likely associated with lithospheric thinning and rifting, occurs across the North Atlantic region in disparate occurrences, typically as small-fraction melts from the Late Triassic to the Cretaceous (Helwig et al., 1974; King and McMillan, 1975; Tappe et al., 2007; Larsen et al., 2009; Peace et al., 2016; Peace et al., 2018c; Peace et al., 2018d). These igneous rocks do not comprise a coherent magmatic province, but rather small-volume, distributed melts (e.g., lamprophyre dykes in West Greenland and Newfoundland; Helwig et al., 1974; Larsen et al., 2009). They demonstrate that significant lithospheric extension was likely widespread across the proto-North Atlantic region as far back as the Late Triassic (Larsen et al., 2009).

During and after breakup, widespread magmatism formed the North Atlantic Igneous Province (NAIP) (White, 1988; Upton, 1988; Saunders et al., 1997; Meyer et al., 2007; Storey et al., 2007; Hansen et al., 2009; Wilkinson et al., 2016; Á Horni et al., 2017). The NAIP is a classic LIP (Bryan and Ernst, 2008; Hansen et al., 2009) that comprises the voluminous Palaeogene igneous rocks of the East Greenland margin (Tegner et al., 1998), NW European margin (Melankholina, 2008), and JMMC (Breivik et al., 2012). To the west of Greenland, in the Davis Strait and on Baffin Island, other Palaeogene igneous rocks contribute to the NAIP (Clarke and Upton, 1971; Upton, 1988; Tegner et al., 2008; Hansen et al., 2009; Gaina et al., 2009; Nelson et al., 2015; Clarke and Beutel, this volume).

Distribution of NAIP volcanism is highly asymmetric between conjugate margins and the more magmatic margins may be associated with thicker lithosphere (Á Horni et al., 2017). Significantly more volcanism occurs south of the GIFR than to the north (Schiffer et al., 2015; Á Horni et al., 2017). Petrologically, NAIP igneous rocks are highly diverse and include tholeiitic and alkali basalts, nepheline- and quartz-syenites, nephelinites, and carbonatites (Holbrook et al., 2001). NAIP igneous rocks are also highly variable in structure and include dykes, and sills (Magee et al., 2014), seaward-dipping reflectors (SDRs) (Larsen and Saunders, 1998), high-velocity lower crustal bodies (Funck et al., 2007), seamounts (Jones et al., 1974), and subaerial flows (Wilkinson et al., 2016; Á Horni et al., 2017).

Although the NAIP is often considered to comprise all pre-, syn- and post-breakup magmas, some are not generally included. For example, the Vestbakken Volcanic Province, and its conjugate equivalent in NE Greenland, have been attributed to local tectonic processes associated with shear margin development and are generally not considered part of the NAIP (Hansen et al., 2009; Á Horni et al., 2017). Significant magmatism is detected by seismic reflection, gravity and magnetic surveys near the western termination of the Charlie-Gibbs Fracture Zone (CGFZ) in the form of multiple flows and seamounts that are not typically considered part of NAIP (Pe-Piper et al., 2013; Keen et al., 2014). The basaltic 'U-reflector' sills offshore Newfoundland, which cover an area of c. 20,000 km^2, are also excluded from the NAIP (Karner and Shillington, 2005; Hart and Blusztajn, 2006; Deemer et al., 2010; Peace et al., 2017b). The logic of inclusion or exclusion of magmatism under the umbrella term NAIP becomes increasingly unclear when it is noted that the Cretaceous-aged Anton Dohrn and Rockall seamounts are considered to belong to NAIP (Hitchen et al., 1995; Morton et al., 1995). This casts doubt on the rationale behind inclusion of igneous rocks in the NAIP and has implications for the extent, timing, magmatic budget and duration of NAIP, which in turn affect models for the tectono-magmatic processes responsible for its development. Much previous work also associates this LIP with a unique geochemical signature, although it is, in fact, highly variable (Korenaga and Kelemen, 2000; Á Horni et al., 2017).

The area of the NAIP has been estimated to be 1.3 × 10^6 km^2, and its volume, which is problematic to assess, is suggested to have once been 5 - 10 × 10^6 km^3 (Holbrook et al., 2001; Storey et al., 2007; Wilkinson et al., 2016). Holbrook et al. (2001) estimated that between breakup and magnetic Chron C23n, 10^7 km^3 of igneous crust was produced. The West Greenland constituent of the NAIP (the West Greenland Volcanic Province; WGVP e.g., Gill et al., 1992) is estimated to cover 2.2 x 10^3 km^2 in area (Clarke and Pedersen, 1976; Riisager et al., 2003).

6.3 Timing of rifting and magmatism

The NAIP is thought to have involved two main periods of melt emplacement: 1) ca. 62-58 Ma and 2) ca. 57-53 Ma, with distinct peaks in productivity at ca. 60 Ma and ca. 55 Ma (Hansen et al., 2009). Distinct parts of the NAIP were emplaced at different times (Lundin and Doré, 2005b). For example the British volcanic province (BVP) and the WGVP are mostly Early Palaeocene whereas NE Atlantic magmatism is predominantly Early Eocene (Lundin and Doré, 2005b). A unifying genetic model must account for this variable spatiotemporal distribution (Lundin and Doré, 2005b; Peace et al., 2017a).

Petersen et al. (2018) recently proposed a mechanism to explain the two-phase igneous activity associated with the NAIP based on numerical modelling. They propose that lithospheric delamination triggered by destabilisation of thickened and metamorphosed, high-density lower crust produced the first igneous peak by small scale convection induced by detachment of the lithosphere. A second, much more voluminous phase of melting occurred when sinking lithospheric blocks penetrated the lower mantle and induced return flow.

In summary, rifting and breakup of the North Atlantic region was accompanied by prolonged, variable and extensive magmatism, some of which is conventionally considered to be part of the NAIP and some of which is not. The distinction is apparently model-dependent, inviting reassessment of both model and categorisation of the magmas.

6.4 Kinematics of the North Atlantic rift – implications for breakup

The North Atlantic opened by south-to-north propagation from the Central Atlantic into the NW and NE Atlantic (Lundin, 2002; Barnett-Moore et al., 2018; Nirrengarten et al., 2018). This contradicts the hypothesis that rifting migrated away from the NAIP, including the WGVP (Lundin and Doré, 2005a; Peace et al., 2017a). On the contrary, rifting migrated towards it, at odds with a plume-driven continental breakup model (Foulger et al. this volume).

There is little evidence for a time-progressive hotspot track (Lundin and Doré, 2005a) as predicted for a plume (Lawver and Müller, 1994; O'Neill et al., 2005; Doubrovine et al., 2012; Mordret, 2018). Although the GIFR is commonly viewed as a plume track there is no seamount chain to support this (Lundin and Doré, 2005b; Foulger et al. this volume). Similarly, in the West Greenland area, Peace et al. (2017a) note that evidence for a distinctive hotspot track associated with the WGVP is vague and poorly constrained, and that rifting and breakup do not follow the predicted path of the proposed plume (Lundin and Doré, 2005a). Additionally, in an idealised plume model, a deep-seated mantle plume would be required to precisely follow lithospheric breakup (e.g., Steinberger et al., 2018) for it to have remained beneath the active spreading plate boundary since inception (Lundin and Doré, 2005b). However, in reality a hypothetical mantle plume may deviate from the idealised model due to a number of processes such as shear flow (Richards and Griffiths, 1988) and deflection around cratonic keels (Sleep et al., 2002).

As described above, extension and magmatism are widely documented prior to postulated plume arrival in the Early Cenozoic. Within the NAIP, the occurrence of significantly more volcanism south of the GIFR than to the north (Schiffer et al., 2015; Á Horni et al., 2017) is at odds with the radial distribution of magmatism predicted by in an idealised plume model.

In summary, breakup of the North Atlantic was a complex, polyphase process, accompanied by highly compositionally variable magmatic events that require numerous ad hoc embellishments of a deep mantle plume impingement model. We it has been suggested that continental breakup and associated magmatism across the North Atlantic region was driven by lithospheric processes associated with plate tectonics (Lundin and Doré, 2005a; Lundin and Doré, 2005b; Ellis and Stoker, 2014; Schiffer et al., 2015; Peace et al., 2017a; Schiffer et al., 2018b), and that mantle temperatures were likely only slightly, if at all, above ambient (Hole and Natland, this volume)

7.0 Discussion

Magmatism is mainly confined to active plate boundaries (i.e., spreading ridges and subduction zones) where plate tectonic processes are indisputably responsible (Kearey et al., 2009). It has been suggested that the same holds true for continental margins.

It was realised early that the dominant force driving plate motion is slab-pull, which is probably an order of magnitude stronger than other forces (Forsyth and Uyeda, 1975). This is consistent with the observation that the speed with which plates move is related to the length of the subducting slab to which they are attached. Considerable work has been done subsequently to investigate this relationship, including study of the apparent east-west asymmetry in the global subduction slab system (Doglioni and Anderson, 2015) and the systematic westward migration of spreading ridges which imparts east-west asymmetry to the composition of the mantle (Chalot-Prat et al., 2017).

In addition, new plate boundaries must, from time to time, be created because the constantly evolving configuration of plates results in periodic annihilation of plate boundaries and transmutation of others (e.g., subduction of the Farallon ridge and replacement of that subduction zone with the San Andreas transform system). Like all cracks in brittle material, extensional plate boundaries are most easily formed by propagation along pre-existing zones of weakness (Holdsworth et al., 2001). The most susceptible zones may well lie in the continental lithosphere, in particular if that lithosphere has been pre-weakened by a long history of tectonic deformation (Butler et al., 1997; Armitage et al., 2010; Audet and Bürgmann, 2011; Petersen and Schiffer, 2016; Peace et al., 2018a). The spatial scaling of lithospheric processes such as rifting and delamination, the heterogeneity of mantle composition (Foulger et al., 2005a; Chalot-Prat et al., 2017) and the complexity of other influential factors such as structural inheritance can explain the great diversity observed along such boundaries (Petersen and Schiffer, 2016; Schiffer et al. this volume). The plate-driven rifting models suggests that continental breakup is initiated by extensional forces, accompanied by rift-shoulder uplift, and magmatism is related to the passive upwelling of local, relatively shallow asthenosphere (Menzies et al., 2002). The extensional forces result from far-field plate-tectonic reorganisations (Geoffroy, 2005).

Plume impingement models predict uplift, LIP-emplacement and rifting in rapid succession (White, 1988; Dam et al., 1998; Beniest et al., 2017; Steinberger et al., 2018). In such models, the bulk of the magmatic products are expected prior to and during the initial stages of rifting, shortly after plume impact. In an ideal, theoretical case, stress in the overriding plate should be concentric around the location of plume impact (Franke, 2013) and lithosphere fragmentation should be initially radial, possibly via multiple rifts, and possibly forming triple junctions (Ernst and Buchan, 1997). Rifting is expected to initiate at, and propagate away from, the point of plume impact and LIP magmatism (Camp and Ross, 2004; Franke, 2013; Peace et al., 2017a). The regions we reviewed, associated with the Pangaea breakup, do not display these features.

7.1 Magmatism associated with Pangaea breakup

Emplacement of the CAMP LIP is the event traditionally associated with plume-driven models for formation of the Central Atlantic (Wilson, 1997). A centre at the Blake Plateau, near the modern-day Bahamas, has been proposed as the focus from which radiating rifts are expected (May, 1971). However, detailed observations do not fit this idealised model (McHone, 2000). Instead of post-dating and emanating from the CAMP LIP, continental rifting preceded it by ~30 Myr, started far to the south and propagated north where rifting continued for 5-10 Myrs after CAMP volcanism ceased (Olsen, 1997). The spatial pattern of volcanism fails to match the predictions. Circum-Atlantic dykes are mostly oriented parallel to adjacent segments of the Central Atlantic rifted margins and a radial model has little support (McHone, 2000). Evidence for a plume track is also lacking since small-volume volcanic features on the Central Atlantic seafloor are much younger than CAMP volcanism, and may be entirely unrelated to CAMP and breakup.

A model for breakup as the culmination of long-term continental tectonic instability (Keppie, 2016), with rifting controlled by reactivation of older structures (Pique and Laville, 1996), and magmas tapped from the asthenosphere, explains the observations more easily (McHone, 2000). The Central Atlantic Ocean opened only after a protracted period of continental rifting (Davison, 2005). The continental margins re-opened sutures that had experienced at least two previous Wilson Cycles of suture and breakup, testifying to the controlling role of inheritance of pre-existing structure (Schiffer et al. this volume). CAMP magmatism comprised a brief phase of ~1 Myr of intense igneous productivity in the midst of a rifting event that lasted several tens of Myr. Volcanic rates were briefly so massive that production cannot be accounted for by any thermal-upwelling mechanism no matter how hot (Cordery et al., 1997). Furthermore, magmas were so widespread, extending throughout a region > 5,000 km wide (Denyszyn et al., 2018) penetrating far into the South American and African continents, that they cannot be attributed to a single source (McHone, 2003; Leleu et al., 2016). Instead, they require widespread lithospheric instability. The petrological diversity of CAMP lavas also cannot be explained by a single source but requires considerable mantle-source heterogeneity, possibly from recycled subducted slabs (Tollo and Gottfried, 1989).

Less information is available from the Western Somali and Mozambique basins which record the breakup of East and West Gondwana (Phethean et al., 2016). More detail needs to be known about the chronological relationships between tectonism and volcanism in the Mwenezi triple junction and Karoo rift and LIP in order to fully test the plume- and plate-driven hypotheses. It is clear, however, that in keeping with observations elsewhere, tectonic unrest was ongoing, with occasional phases of inactivity, in the region since the Early Permian, over 100 Myr before Jurassic breakup (Macgregor, 2018). Thus, the structures along which breakup-related magmatism occurred predated breakup by many millions of years. For example, many of the dykes in the Okavango swarm were formed in the Proterozoic, and share geochemical affinities with the Mesozoic breakup-related intrusives. This suggests a long-lived volcanic lithospheric feature and source since the region must have moved relative to the deeper mantle in the interim period. Evidence for extensive lateral flow of magmas at the time of breakup testifies to distributed sources rather than a single centre, e.g., at the Mwenezi triple junction. Furthermore, breakup and formation of the ocean basins did not radiate from the Mwenezi triple junction. Instead the evidence available suggests instead that breakup-related rifting migrated towards the triple junction. The close proximity of volcanic margins with SDRs and magma-poor margins is incompatible with a single, large-scale source.

Considerably more is known about the opening of the South Atlantic and the chronology and composition of lavas of the Parana-Etendeka LIP (Foulger, 2017). This region is associated with the Cretaceous disintegration of West Gondwana and it exhibits extensive volcanic margins and SDRs (Franke et al., 2010). In plume models, the large, well-studied Paraná–Etendeka LIP in Brazil and Namibia is attributed to the head of a plume currently beneath Tristan da Cunha (Peate, 1997).

Since the proposal that South Atlantic breakup was plume-driven, a great deal of new and detailed information has accumulated from numerous marine geophysical experiments (Franke et al., 2007; Franke et al., 2010; Foulger, 2017). In addition, the structure and geochemistry of Paraná–Etendeka LIP lavas and postulated 'plume tail' volcanics on the Rio Grande- and Walvis ridges have been critically examined. Major chronological and spatial mismatches with the plume-driven breakup model have emerged. Rifting onset occurred long before the Paraná–Etendeka LIP was emplaced at ~132 Ma. Seafloor spreading in the southern South Atlantic in the Valanginian, at ~135 Ma and propagated northward in jumps, with brief hiatuses where the developing rift encountered barriers. Major volcanic margins were built, and thus breakup and large-scale magmatism was already well underway when the Paraná–Etendeka LIP was emplaced, at odds with the plume-driven breakup model. The rift unambiguously propagated toward the future location of the LIP, not away from it (Foulger, 2017).

Paraná lavas were emplaced in north-south-trending rift basins, testifying to ongoing extension prior to LIP emplacement. They erupted at the location of a major cross-cutting transverse lineament (Foulger, 2017), exploiting pre-existing structure. Of all continental LIPs, the geochemistry of the Paraná–Etendeka LIP is also perhaps the least equivocal that the lavas were derived from melted lithospheric mantle. In addition, recent detailed seismic surveys, both of the breakup margins and the African coastal part of the Walvis Ridge, show that spatially abrupt changes in magma volume are widespread (Franke et al., 2010).

The complex history of North Atlantic breakup and magmatism has been studied intensely for many decades, and is known in detail (Clarke and Upton, 1971; Srivastava and Roest, 1999; Hansen et al., 2009; Larsen et al., 2009; Nirrengarten et al., 2018; Schiffer et al. this volume; Hole and Natland, this volume). Volcanism has been widespread since the region initially began rifting in the Early Jurassic (or possibly Late Triassic; Larsen et al., 2009) followed by opening of the Labrador Sea (Chalmers and Pulvertaft, 2001). Early, relatively small-volume volcanism (Peace et al., 2018c) gave way to emplacement of massive volcanic margins with SDRs when spreading was transferred to the current NE Atlantic (Eldholm and Grue, 1994; Wilkinson et al., 2016).

Several magmatic events have been attributed to an Icelandic plume head, including the Siberian Traps (~251 Ma), volcanism in the Davis Strait (~62 Ma) (Gerlings et al., 2009) and widespread magmatism at the time of opening of the NE Atlantic Ocean (~54 Ma) (Steinberger et al., 2018). The latter two events accompanied lithospheric breakup and there is no evidence of a chronology of uplift followed by LIP volcanism and subsequent continental rifting (Foulger and Anderson, 2005; Peace et al., 2017a). On the contrary, tectonic unrest, continental extension and small-volume volcanism for several 100 Myr prior to breakup is well-documented in Laurasia prior to breakup (Tappe et al., 2007; Larsen et al., 2009; Peace et al., 2016; Peace et al., 2018c).

Continental breakup along the Labrador Sea axis propagated to the region from the south (Chalmers and Pulvertaft, 2001; Peace et al., 2018a), and considerable magmatism occurred prior to emplacement of the magmas usually attributed to a plume head (McWhae et al., 1980; Larsen et al., 2009). At the Davis Strait and the GIFR, that magmatism occurred at locations where propagating breakup rifts encountered barriers that stalled progress (Peace et al., 2017a).

In the case of the GIFR, a major focus of plume models (Foulger et al. this volume), volcanism developed at a locality where rifts propagating from both north and south were unable to break through transverse inherited orogenic structures.

7.2 Summary of spatial-temporal and magmatic-lithospheric relationships

All the locations reviewed herein show evidence for prolonged phases of rifting prior to LIP magmatism and breakup. In many cases, this rifting is thought to be genetically linked to breakup (e.g. rifting prior to the opening of the North Atlantic; e.g., Péron-Pinvidic et al., 2017). At other locations earlier rifting events may have been unassociated with the final breakup episode and the production of the first true oceanic crust. In all cases reviewed here, the onset of LIP magmatism and often eruption of the main volume significantly overlapped with or postdated SDR- and initial-oceanic-crust production. Such magmatism is inconsistent with plume impact driving sometimes long-lasting initial rifting. Instead, it suggests that magmatism was a consequence of the same mechanism that triggered by rifting and/or breakup.

Following plume arrival, widespread magmatism is predicted to occur in the region underlain by hot plume head material (Saunders et al., 1992; Saunders et al., 2007). This region is inferred to be circular, with a diameter of several 1000 kilometres in an idealised model. Buoyant melt is expected to intrude the crust radially, governed by the circular stress field generated by the impinging plume, and to form radial dyke swarms and sills, again in an idealised model. Lithospheric structure is expected to impose only secondary control (Saunders et al., 2007). The relatively small barriers presented by lithospheric inhomogeneities are expected to be overwhelmed by the much larger scale hot upwelling mantle material. These predictions are, however, not supported by observations of the disintegration of Pangaea. Instead, inherited lithospheric structure exerts a control, not only on the locus of breakup axes but also on the locations of magmatism including LIPs (Koopmann et al., 2014a; Peace et al., 2017a; Clarke and Beutel, this volume).

7.3 Plate-driven breakup

Plate-driven breakup models for the dispersal of Pangaea have been proposed. For example Keppie (2016) proposed that subduction at the peripheries of Pangaea can explain both the motion, deformation and dispersal of Pangaea with a single mechanism. In addition, much previous work links continental rifts and breakup on a range of scales and tectonic environments to pre-existing structures (Wu et al., 2016; Petersen and Schiffer, 2016; Peace et al., 2018b; Schiffer et al., 2018; Collanega et al., 2019; Schiffer et al., this volume). A link between the intersection of propagating rifts with pre-existing suture zones and the production of magmatism has been suggested based primarily on geological observations from Atlantic margins and numerical modelling (Koopmann et al., 2014a; Schiffer et al., 2015; Peace et al., 2017a; Petersen et al., 2018). In these models, a barrier to rift propagation results in excess magmatism by blocking and diverting mantle flow beneath a propagating rift axis.

Observations from the locations reviewed here provide support for this model (Fig. 6). In the North, Central and South Atlantic and during breakup of East and West Gondwana, LIP locations coincide with large-scale, pre-existing lithospheric structures (Fig. 7). The origin, size and relative orientations of these structures with respect to approaching, propagating rifts is variable. Nevertheless, the association is systematic and warrants further investigation.

7.4 Ocean Island Chains

The mantle plume hypothesis for LIP volcanism predicts that, following plume-head-related flood-basalt eruptions, continued upwelling in the plume tail results in ongoing, small-volume magmatism (Saunders et al., 1992; White, 1992). The motion of the overhead plates relative to the "hotspot reference frame" transports these magmas away from the plume tail creating a time-progressive trail of volcanism that ages with increasing distance from the contemporary plume tail (e.g., Konrad et al., 2018). The existence of a time-progressive trail of volcanism, most clearly observed in the Hawaiian-Emperor island/seamount chains, was the single most influential factor in the development of the plume hypothesis and this characteristic is still considered by some to comprise the strongest evidence of a mantle plume (Morgan and Morgan, 2007).

This aspect of the plume model fits poorly the volcanism that followed emplacement of the LIPs discussed in this paper. Courtillot et al. (2003) review the features of postulated plumes worldwide. Of the four volcanic provinces we discuss (the CAMP, Karoo-Ferrar flood basalts, South Atlantic Igneous Province and NAIP), Courtillot et al. (2003) associate only the South Atlantic Igneous Province LIP flood-volcanism unambiguously with a time-progressive volcanic trail. Moreover, Courtillot et al. (2003) tentatively associate the CAMP with small volumes of volcanism at Fernando de Noronha on the Brazilian continental shelf and minor volcanism onshore. Considering the CAMP is thought to be one of the largest continental LIPs in the world (Denyszyn et al., 2018), the minimal evidence for plume tail volcanism makes the model doubtful. In addition, conflicting evidence from geochemistry (Lopes and Ulbrich, 2015) and the chronology of volcanism along archipelago of Fernando de Noronha (Knesel et al., 2011) casts further doubt on the applicability of an idealised plume model. The Karoo-Ferrar flood basalts are tentatively associated with a postulated plume currently centred beneath Crozet/Prince Edward Island (Courtillot et al., 2003). The volcanics that represent the best candidates for a time-progressive trail extending from that region comprise a ~200-km-wide archipelago of five island groups across which recent volcanism is widespread and evidence for systematic time-progression sparse. However, the oldest volcanism known is ~9 Ma (Verwoerd et al., 1990)and there is no apparent link with the ~185-177 Ma Karoo-Ferrar flood basalts.

There is evidence for some age progression in volcanics in the South Atlantic. Proposed plume-tail volcanism comprises the Rio Grande Rise, the Walvis aseismic ridge and the associated Guyot Province that extends from the Etendeka continental flood basalts in Namibia to the volcanically active island of Tristan da Cunha. Reported ages range from 114 Ma near Namibia to 58–72 Ma at the SW end of the Walvis Ridge, and to 80–87 Ma for the Rio Grande Rise, which is believed to represent the counterpart of the Walvis Ridge on the South American Plate (Rohde et al., 2013). Age-progressive dates are obtained from the Walvis Ridge (O'Connor and Jokat, 2015) but there is little corresponding evidence from the Rio Grande Rise. On the contrary, continental material has recently been observed there, suggesting that the Rise is possibly a micro-continental fragment (Sager, 2014) that could have been isolated by a series of eastward ridge jumps (Graça et al., 2019).

In the NAIP the voluminous flood volcanism that formed the North Atlantic passive margins, is popularly attributed to a plume head (e.g., Chalmers et al., 1995; Gill et al., 1995; Steinberger et al., 2018). However, it is associated with no observed time-progressive volcanic trail (Peace et al., 2017a; Foulger et al. this volume). The GIFR is often attributed to this but supporting evidence is lacking (Foulger et al. this volume). Very few seamounts occur on the GIFR (Gaina et al., 2017a) and few reliable dates are available. The GIFR is time-progressive only in the same sense as the ocean floor, and it is interpreted to have formed as a consequence of prolonged, highly volcanic lithospheric extension (Foulger et al. this volume).

8.0 Concluding remarks

This review highlights significant spatial-temporal variability between the locations of LIPs and the initiation points of Pangaea disintegration. None of the regions we review fit comfortably a plume-driven breakup model that predicts pre-breakup magmatism, plume tail eruptions producing ocean island chains, and rifting radiating from the point of plume impact. In contrast, most show multiple characteristics that are not fully compatible with this model, including a reverse chronology of uplift, magmatism and rifting, and rifting propagating towards LIPs. The idealised, generic plume-impingement model thus has difficulties fully explaining the dispersal of Pangaea and associated magmatism.

Rifting and breakup driven primarily by far-field extensional forces, with magmatism occurring as a consequence, under strong lithospheric control, is much more consistent with observations that are common throughout the regions we review. These observations include:

- The supercontinent in the neighbourhood of future breakup experienced almost continuous unrest, including extension and continental rifting and small-volume magmatism for long periods prior to breakup (10s to 100s of Myr).
- Evidence for pre-LIP uplift is lacking. Margin uplift contemporaneous with breakup is consistent with rift-shoulder uplift.
- Magmatism followed pre-existing structures that may have experienced volcanism before.
- The source of magmas was distributed. Magmas did not arise from a single centre.
- Large-volume magmatism (LIP emplacement) occurred distal to simultaneous breakup-related rifting, which tended to migrate towards the new LIP.
- The geochemistry of LIP lavas, in particular their Ti contents, suggest a source in the lithospheric mantle.
- The very rapid emplacement of the LIP lavas, with rates on the order of 10^6 km^3 in 1 Myr, are incompatible with melt production on the same time-scale as eruption. They can essentially only be explained as the draining of pre-existing melt reservoirs that accumulated over a longer period of time than it took to drain them (Silver et al., 2006).

Other factors that likely exert some influence include spatially and temporally variable mantle-source temperature and composition as rifts propagate laterally and asthenosphere wells up from beneath lithosphere initially 100-200 km thick (Brandl et al., 2013; Langmuir, 2013). Other processes that may encourage or be consequential to rifting include delamination of lower lithosphere, small-scale convection (King and Anderson, 1995; King and Anderson, 1998; Simon et al., 2009; Peace et al., 2017a) along Archean craton boundaries and fragmentation of the new margins to form microcontinents (Schiffer et al., 2018). In conclusion, a lithosphere-centred model for Pangaea breakup is the simplest that can explain the primary, common features expressed along the passive margins of the former supercontinent Pangaea.

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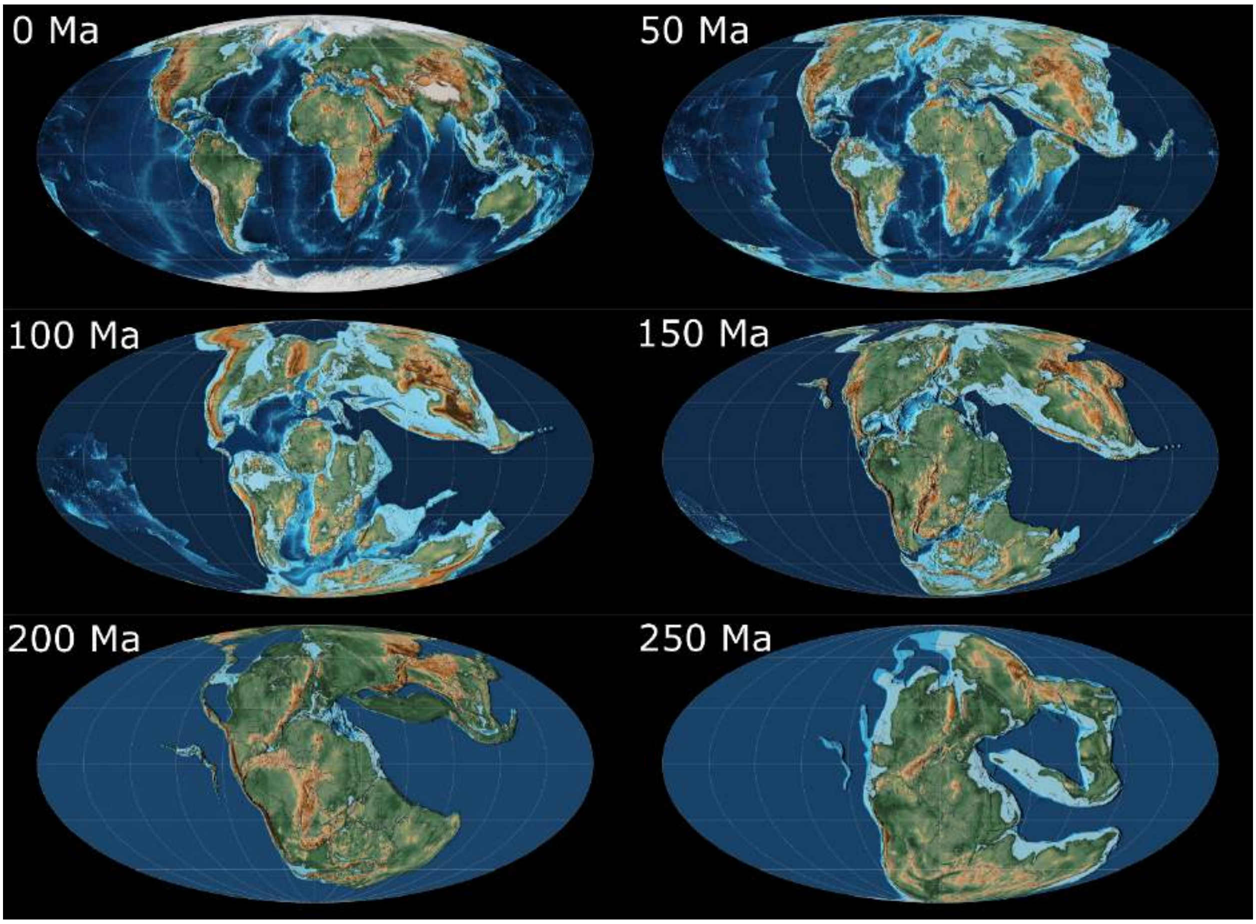

Figure 1. An overview of the disintegration of Pangaea (e.g., Frizon De Lamotte et al., 2015) using the palaeogeographic reconstruction compiled into the PALEOMAP PaleoAtlas for GPlates (Scotese, 2016) *plotted using a Mollweide projection and shown at 0, 50, 100, 150, 200 and 250 Ma.*

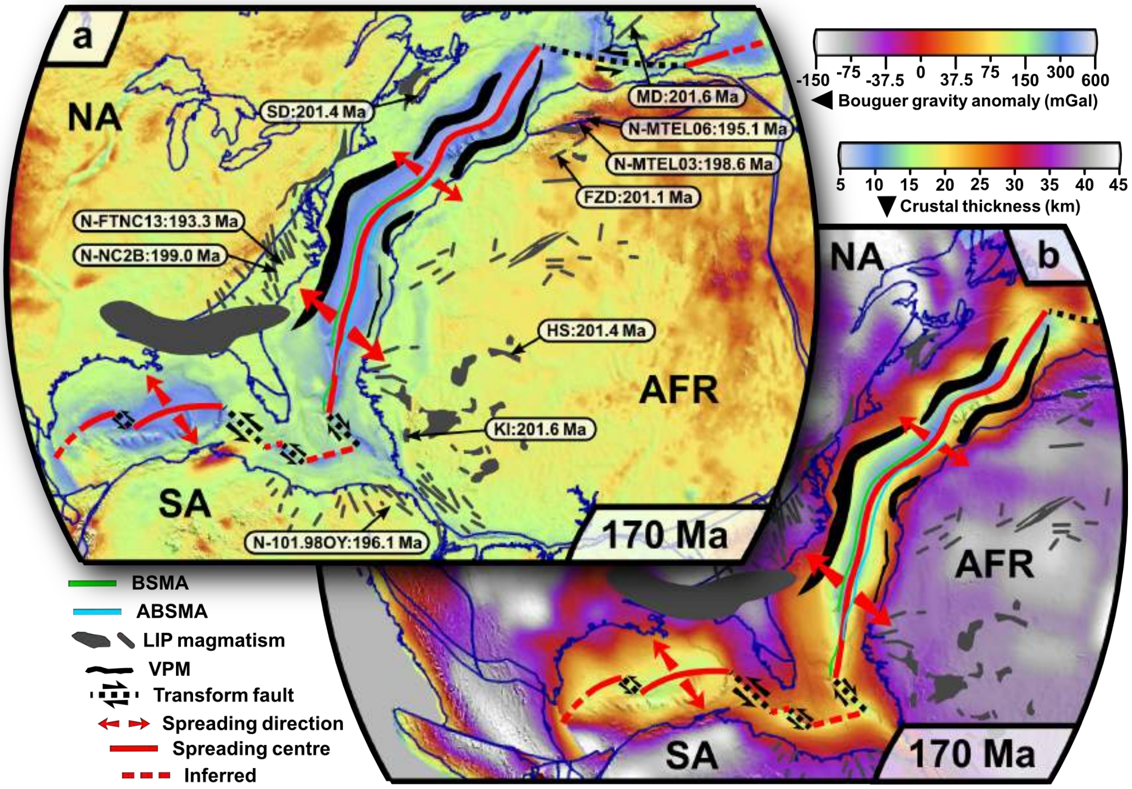

Figure 2. Breakup of the Central Atlantic shown at 170 Ma. a) Reconstructed present day Bouguer gravity anomaly (world gravity map; Balmino et al., 2012). b) Reconstructed present day crustal thickness according to the CRUST1.0 model (Laske et al., 2013). *Representative LIP magmatism, SDRs, and earliest oceanic magnetic anomalies are shown with associated ages where available. NA = North America, SA = South America, AFR = Africa.*

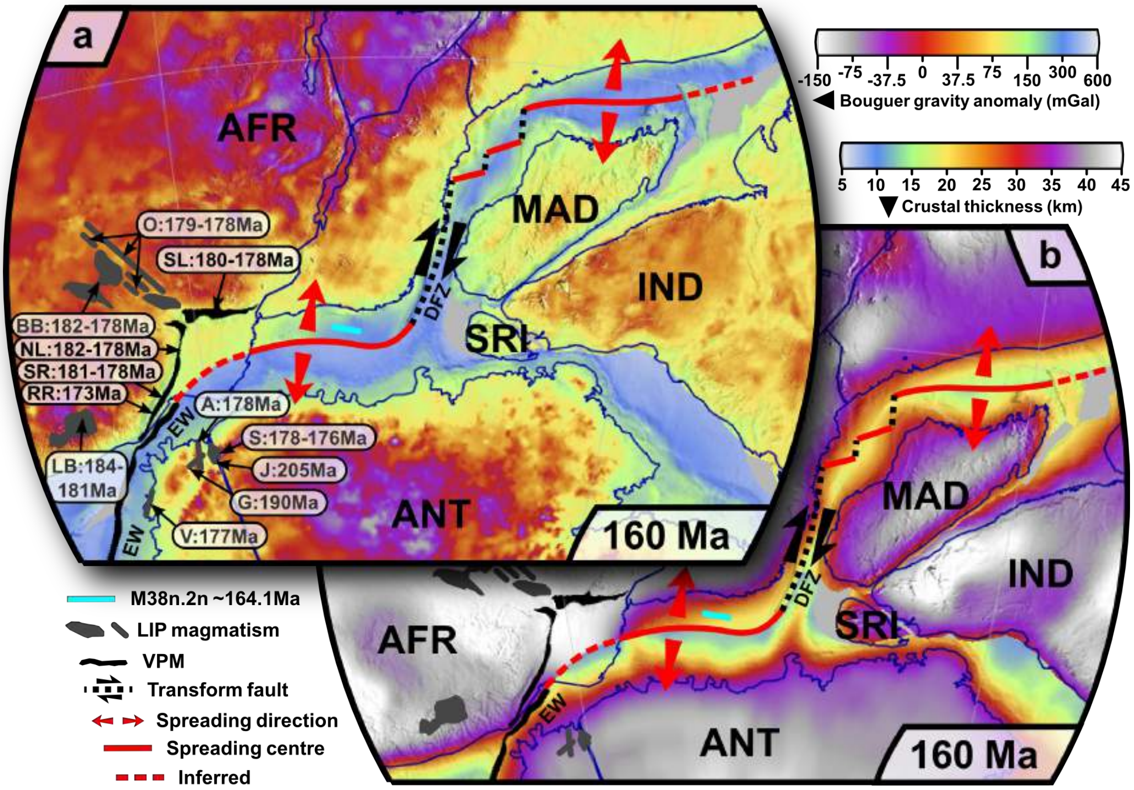

Figure 3. Breakup of East and West Gondwana shown at 160 Ma. a) Reconstructed present day Bouguer gravity anomaly (world gravity map; Balmino et al., 2012). b) Reconstructed present day crustal thickness according to the CRUST1.0 model (Laske et al., 2013). Representative LIP magmatism, SDRs, and earliest oceanic magnetic anomalies are shown with associated ages where available (Phethean et al., 2016; Klimke and Franke, 2016; Sauter et al., 2018). AFR = Africa, MAD = Madagascar, DFZ = Davie Fracture Zone; IND = India, SRI = Sri Lanka, ANT = Antarctica, and EW = Explora Wedge.

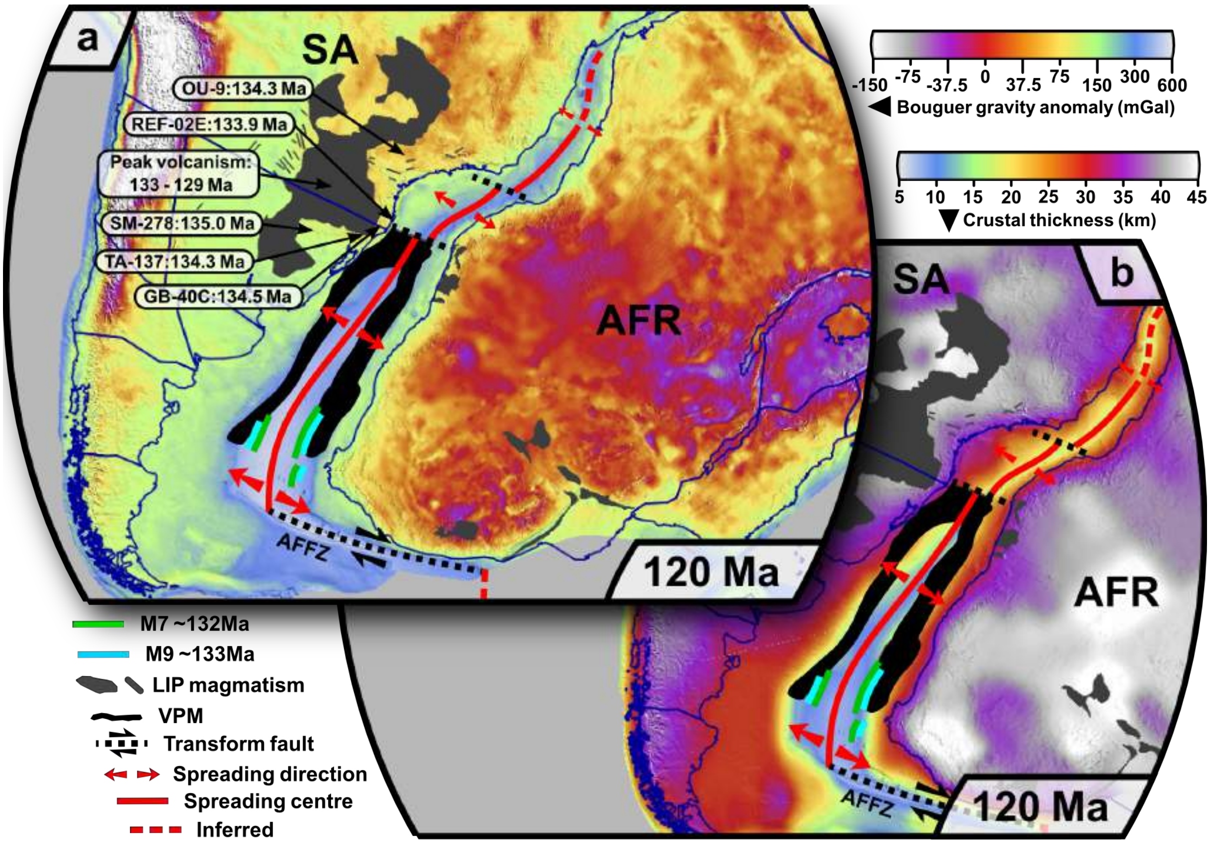

Figure 4. Breakup of the Southern Atlantic shown at 120 Ma. a) Reconstructed present day Bouguer gravity anomaly (world gravity map; Balmino et al., 2012). b) Reconstructed present day crustal thickness according to the CRUST1.0 model (Laske et al., 2013). *Representative LIP magmatism, SDRs, and earliest oceanic magnetic anomalies are shown with associated ages where available (Koopmann et al., 2016). AFR = Africa, SA = South America, & AFFZ = Agulhas-Falkland Fracture Zone.*

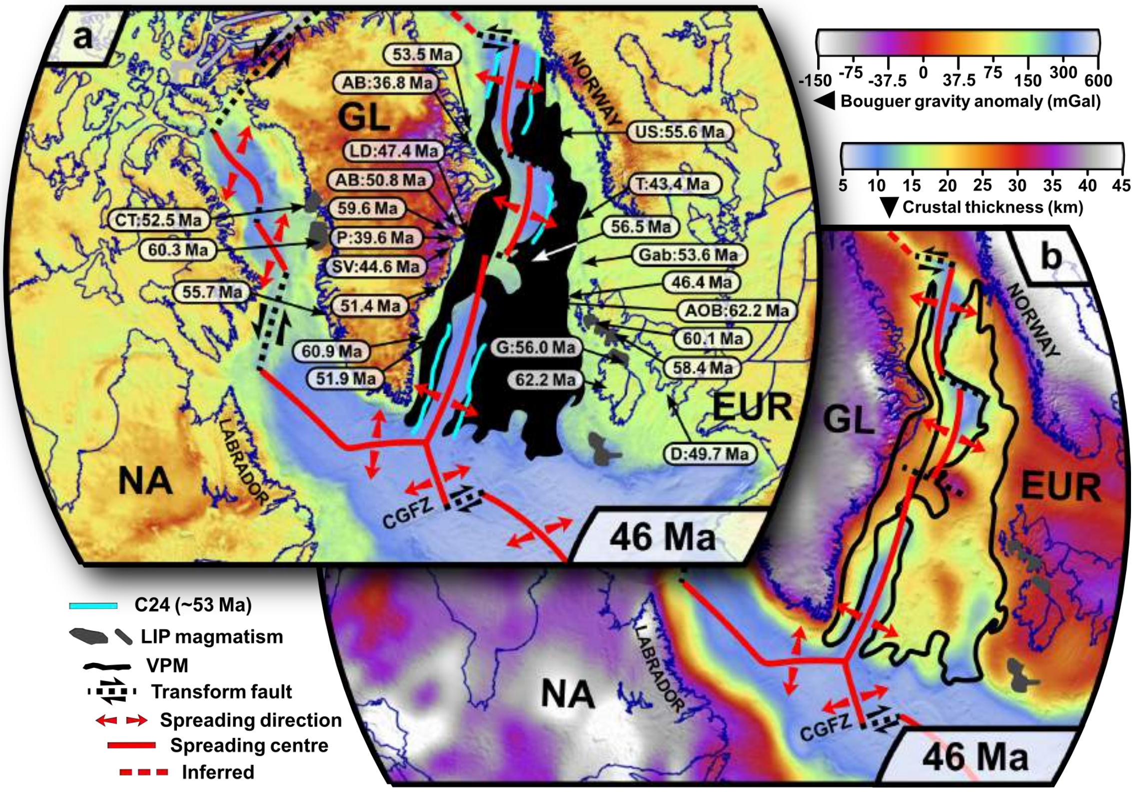

Figure 5. Breakup of the North Atlantic shown at 46 Ma. a) Reconstructed present day Bouguer gravity anomaly (world gravity map; Balmino et al., 2012). b) Reconstructed present day crustal thickness according to the CRUST1.0 model (Laske et al., 2013). LIP magmatism, SDRs, and earliest oceanic magnetic anomalies are shown with associated ages where available. Representative magmatism ages are primarily modified from the compilation made for the NAGTEC project (Wilkinson et al., 2016). Location names: AB = Alkaline Basalt; US: NA = North America, GL = Greenland, EUR = Europe, and CGFZ = Charlie-Gibbs Fracture Zone.

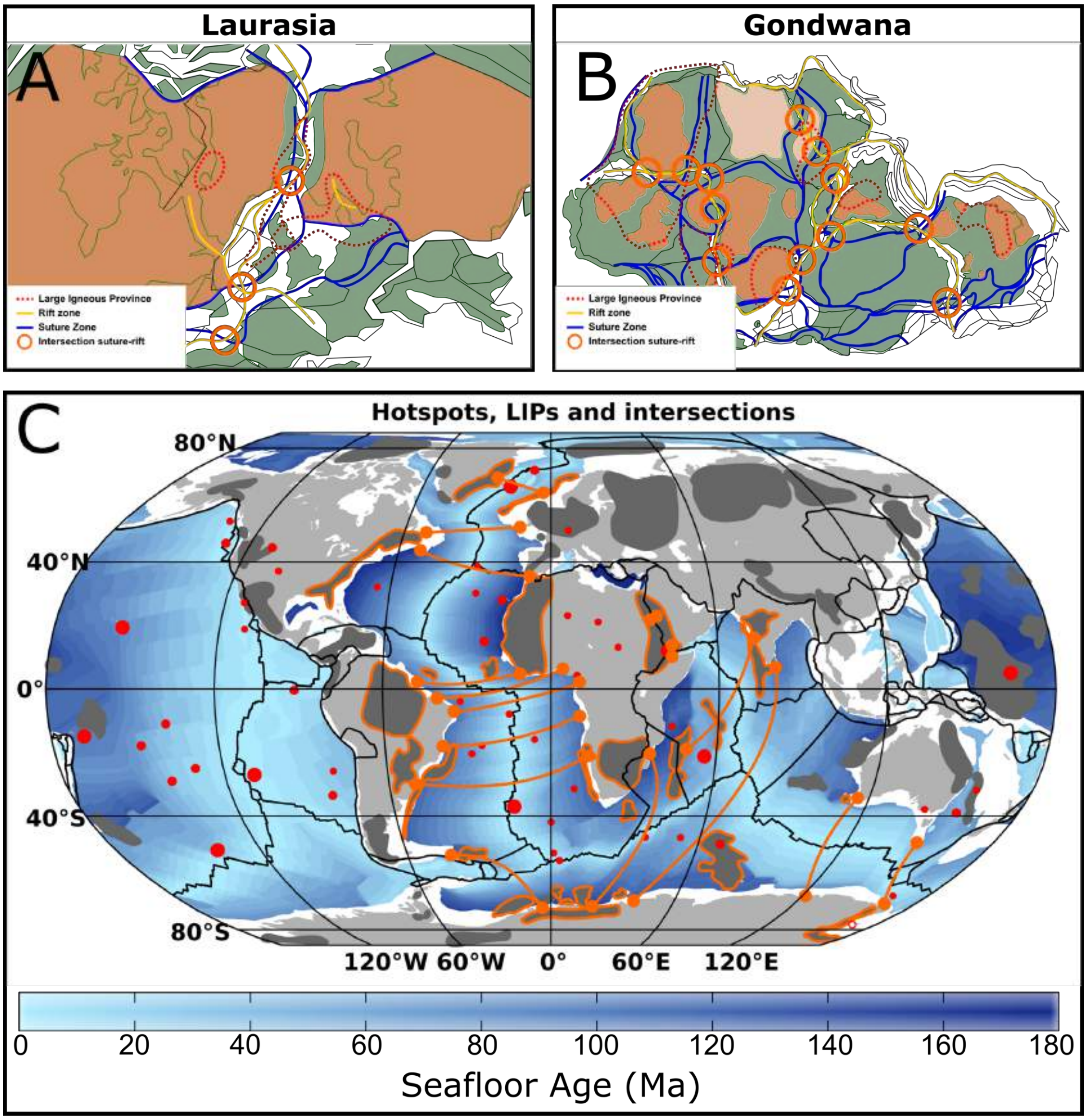

Fig. 6. Schematic reconstructions of A) Laurasia (Cocks and Torsvik, 2011) *and B) Gondwana (Stampfli et al., 2013) where: green = present-day land areas; brown = cratons; blue lines = suture zones; yellow = incipient breakup axes; orange circles = intersection of breakup axes with suture zones and red dotted lines = schematic outline of LIPs. C) A global overview of the relationship between continental crust (white=offshore; pale grey = onshore), LIPs (dark grey), proposed hotspots (red dots) and the reconstructed pre-rift intersection points between suture zones and continental breakup (orange dots). Orange borders on LIPs indicate those that may have been involved with Pangaean dispersal. The size of the red dots (representing hotspots) is related to their depths proposed by Courtillot et al. (2003) such that large dots = core-mantle boundary; medium dots = the base of the upper mantle; and small dots = the lithosphere. The orange lines show the interpolation between conjugate intersection points, and the age of oceanic crust is shown in blue. This figure illustrates the relationship between breakup-suture intersections and many LIPs that formed between the conjugate margins where intersection points existed. LIPs on this figure are taken from Ernst (2014). Seafloor age is from Seton et al. (2012).*