Tectonic and magmatic segmentation of the Global Ocean Ridge System: a synthesis of observations

SUZANNE M. CARBOTTE¹*, DEBORAH K. SMITH², MATHILDE CANNAT³ & EMILY M. KLEIN⁴

¹*Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY 10964, USA*

²*Woods Hole Oceanographic Institution, Woods Hole Massachusetts 02543, USA*

³*Equipe de Géosciences Marines, CNRS-UMR 7154, Institut de Physique du Globe, 1 rue Jussieu, 75238 Paris cedex 05, France*

⁴*Nicholas School of the Environment, Duke University, Durham, NC 27708, USA*

*Corresponding author (e-mail: carbotte@ldeo.columbia.edu)

Abstract: Mid-ocean ridges display tectonic segmentation defined by discontinuities of the axial zone, and geophysical and geochemical observations suggest segmentation of the underlying magmatic plumbing system. Here, observations of tectonic and magmatic segmentation at ridges spreading from fast to ultraslow rates are reviewed in light of influential concepts of ridge segmentation, including the notion of hierarchical segmentation, spreading cells and centralized v. multiple supply of mantle melts. The observations support the concept of quasi-regularly spaced principal magmatic segments, which are 30–50 km long on average at fast- to slow-spreading ridges and fed by melt accumulations in the shallow asthenosphere. Changes in ridge properties approaching or crossing transform faults are often comparable with those observed at smaller offsets, and even very small discontinuities can be major boundaries in ridge properties. Thus, hierarchical segmentation models that suggest large-scale transform fault-bounded segmentation arises from deeper level processes in the asthenosphere than the finer-scale segmentation are not generally supported. The boundaries between some but not all principal magmatic segments defined by ridge axis geophysical properties coincide with geochemical boundaries reflecting changes in source composition or melting processes. Where geochemical boundaries occur, they can coincide with discontinuities of a wide range of scales.

Discovery and interpretations of mid-ocean ridge (MOR) segmentation: historical review

Early maps of the global mid-ocean ridge (MOR) system showed that spreading centres are segmented at scales of up to hundreds of kilometres by large-offset transform faults (TFs), some of which are inherited from continental breakup. As more detailed maps of the ocean floor became available in the 1970s and 1980s, ubiquitous smaller offset discontinuities were discovered that segment the ridge axis at shorter intervals (e.g. Hey 1977; Hey *et al.* 1980; Schouten & Klitgord 1982; Lonsdale 1983, 1985; Francheteau & Ballard 1983; Macdonald *et al.* 1984; Schouten *et al.* 1985; Langmuir *et al.* 1986; Sempéré *et al.* 1990). These discontinuities were found to be distinct from TFs because they typically lack the narrow zone of spreading-parallel lineaments indicative of localized strike-slip faulting. Together, TFs and these non-transform discontinuities define the tectonic segmentation of the global MOR system.

An important type of non-transform offset (NTO), first recognized at intermediate-spreading ridges, is called a propagating rift (PR) because, at these discontinuities, one ridge segment, the 'PR', advances into older crust adjacent to the 'dying rift', which retreats (Hey 1977; Hey *et al.* 1980). PRs leave a distinctive V-shaped wake of magnetic and bathymetric anomalies on the ridge flanks. Along the fast-spreading East Pacific Rise (EPR), an NTO, similar in many ways to a PR, was later discovered, called an overlapping spreading centre (OSC), where two *en echelon* ridge segments overlap (e.g. Macdonald & Fox 1983; Lonsdale 1983; Macdonald *et al.* 1984, 1988). Like PRs, OSCs can persist for millions of years and, in most cases, migrate along the ridge axis, leaving V-shaped wakes of disrupted seafloor. On an even finer scale, high-resolution mapping of the EPR (e.g. Macdonald *et al.* 1984) led to the identification of yet smaller, shorter-lived ridge offsets, referred to as 'devals' for deviations from axial linearity (Langmuir *et al.* 1986).

At slow-spreading ridges, 'zero-offset transform faults (ZOTs)' were identified from small offsets in

magnetic anomalies in old North Atlantic crust, which were inferred to persist for tens of million years while also changing in length and sense of offset (e.g. Schouten & White 1980; Schouten & Klitgord 1982). Subsequent mapping investigations of the Mid-Atlantic Ridge (MAR) axis revealed small jogs or oblique-trending portions of the rift valley spaced at similar intervals as the ZOTs inferred in old crust (e.g. Schouten *et al.* 1985; Sempéré *et al.* 1990). In some places these NTOs remain stationary and in other places they migrate along the ridge axis (e.g. Schouten *et al.* 1987).

Another early discovery was that regular variations in the depth of the ridge axis and other along-strike ridge properties accompany the tectonic segmentation (e.g. Le Douaran & Francheteau 1981; Francheteau & Ballard 1983; Macdonald *et al.* 1984; Schouten *et al.* 1985). TFs and the newly recognized smaller offset NTOs were found to coincide with local deep points along the ridge axis, with the ridge shoaling away from these discontinuities, forming arched-shaped along-axis topographic profiles. These variations in seafloor depth were attributed to changes in the density structure of the crust and mantle below, with the deep ridge regions interpreted as sites of higher densities resulting from thinner crust and/or cooler crust and upper mantle temperatures (e.g. Francheteau & Ballard 1983; Macdonald *et al.* 1984). While initial studies attributed the deepening of the ridge axis near TFs to a decrease in asthenospheric melting due to the juxtaposition of older and hence colder lithosphere across the discontinuity (e.g. Fox & Gallo 1984), further work showed that the predicted conductive cooling effects were insufficient to generate the increase in seafloor depth (e.g. Forsyth & Wilson 1984). Furthermore at OSCs and smaller discontinuities along the EPR, thermal edge effects due to the contrast in lithospheric age across these offsets (<0.25 Ma) are expected to be negligible. An alternate explanation for the apparent decrease in melt volume near these discontinuities was therefore needed.

The concept of a 'spreading cell' was introduced in the early 1980s to explain axial variations associated with tectonic segmentation. Based on the apparent long-lived coherence of oceanic lithosphere accreted between fracture zones in the North Atlantic and what were later recognized as NTOs along the MAR, and the quasi-regular spacing of these ridge segments, Schouten & White (1980), Schouten & Klitgord (1982) and Schouten *et al.* (1985) advanced the notion of a 'spreading centre cell', a length of ridge representing an independent, long-lived, accretionary unit. Drawing on data from the EPR, Francheteau & Ballard (1983) further suggested that these spreading cells exhibit systematic along-axis topographic, volcanic, tectonic and

hydrothermal gradients, which may result from central supply of melt to segment centres and redistribution of melt toward segment ends. In this model, the deeper seafloor typically found at NTOs, which are too small in offset to produce significant cold edge effects, reflects diminished magma supply at the distal ends of the upwelling centres (Macdonald *et al.* 1984). These and subsequent studies led to the view that each spreading cell has its own physically distinct melt supply system in the asthenosphere; and, further, that the relatively regular spacing of these melt supply systems results from diapiric ascent of Rayleigh–Taylor-type gravitational instabilities that develop in partially molten mantle (Schouten & Klitgord 1982; Francheteau & Ballard 1983; Whitehead *et al.* 1984; Crane 1985; Schouten *et al.* 1985; Rabinowicz *et al.* 1987).

With increasing recognition that the MOR system exhibits segmentation at a range of spatial scales, the model of melt supply centres evolved to include the possibility of a hierarchy of nested magmatic segmentation with a branching, along-axis melt redistribution system arising from variations in the depth and spatial scales of mantle upwelling, melting, melt segregation and delivery (Fig. 1; Langmuir *et al.* 1986; Macdonald *et al.* 1988). Geochemical studies of along-axis variations in melt composition played a key role in exploring this model, based on the assumption that geochemical segmentation (lengths of ridge with common geochemical characteristics) is representative of magmatic segmentation (lengths of ridge fed by a common magma plumbing system). Within the larger-scale segments identified as spreading cells, geochemical studies identified along-axis variations in magma composition reflecting differences in mantle source composition and extents and pressures of melting at shorter spatial scales, suggesting multiple melt injection sites beneath these segments (e.g. Langmuir & Bender 1984; Langmuir *et al.* 1986; Niu & Batiza 1994; Reynolds & Langmuir 1997). Detailed sampling of the fast-spreading EPR revealed that even the smallest tectonic boundaries (devals) can coincide with changes in magma compositions (Langmuir *et al.* 1986).

Other models have been proposed for the origin of ridge segmentation and can be grouped broadly into those that invoke tectonic processes associated with the rifting of young lithosphere (e.g. Sandwell 1986; Lonsdale 1989) and those that invoke mantle upwelling, melting processes, melt segregation and transport (e.g. Whitehead *et al.* 1984; Crane 1985; Schouten *et al.* 1985; Rabinowicz *et al.* 1987; Spiegelman & McKenzie 1987; Parmentier & Phipps Morgan 1990; Sparks & Parmentier 1991, 1993; Lin & Phipps Morgan 1992; Nicolas *et al.* 1994; Rabinowicz & Briais 2002; Toomey *et al.* 2007; Hebert & Montési 2011). The prevailing

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(a) 1st-3rd Order Discontinuities and Segments

(b) 4th Order Discontinuities and Segments

Fig. 1. Schematic illustration of hierarchical central magma supply model for magmatic segmentation from Macdonald *et al.* (1988). (a) Axial depth profile shows undulations in seafloor depth along the ridge axis with discontinuities of orders 1–3 located at local depth maximum. Transform faults or first-order discontinuities are located at the deepest points along the ridge; second- and third-order offsets are associated with smaller depth anomalies; these discontinuities bound segments of different orders (as indicated with labelled horizontal arrows above). Asthenospheric upwelling at depth is enhanced beneath the shallowest part of the ridge. In the shallower mantle, melt segregation events lead to focusing of magma supply beneath the mid-segment regions of second- and third-order segments, followed by along-axis redistribution of magma. (b) Focused mantle melt supply replenishes crustal magma chambers and leads to the finest fourth-order scale of segmentation. Note that discontinuity and segmentation number orders are related; a segment of any order is bound by a discontinuity of the same order at one but not necessarily both ends, with the highest-order offset number governing the designation of the segment order number (e.g. a segment bounded by first- and third-order offsets is a third-order segment).

view is that segmentation at the larger scales arises primarily from the thermochemical properties of the mantle that control patterns of mantle upwelling

and melting. However, the dominant mechanisms are poorly understood; multiple processes may contribute at different ridges, perhaps at different length

and depth scales, and feedbacks between lithospheric and mantle processes are likely important.

The focus of this review

From the numerous studies conducted in the *c.* 30 years since the discovery of MOR segmentation, much has been learned concerning the relationship between tectonic and magmatic segmentation. The identification of tectonic segments, especially those defined by small discontinuities, has been further refined with improved seafloor mapping capabilities. The characterization of magmatic segmentation has also improved with higher resolution geophysical imaging of melt distribution beneath the seafloor and more extensive sampling and chemical characterization of ocean ridge basalts. While the initial models for ridge segmentation were derived primarily from observations of the fast-spreading EPR and slow-spreading MAR, numerous studies of intermediate and ultra-slow ridges have now been conducted that facilitate comparisons of the style and characteristics of segmentation across the full spectrum of spreading rates. Furthermore, dense sampling and focused observations of a range of ridge properties are now available for several regions encompassing different scales of tectonic segmentation suitable for assessing the relationships with magmatic segmentation in detail.

In this contribution we review current observations on MOR segmentation to address the overarching question: Are the different scales of seafloor tectonic segmentation of the MOR coincident with magmatic segmentation of the ridge and, if so, what processes in the crust and mantle magmatic systems are involved?

Drawing upon select regions of ridges spreading at rates from fast to ultra-slow, we review relevant geological, geophysical and geochemical observations. Of particular interest are implications of these observations for the notion of spreading cells, models of central supply v. multiple supply sites beneath segments and the hierarchy of segmentation. Our goal is to provide a summary of current observations for readers interested in MOR studies and magmatic rifting more generally. Given the broad scope of this contribution and extensive literature on the subject, we highlight references that are pivotal to the development of particular ideas. For further discussion of ideas on the origin of segmentation, the reader is referred to excellent prior reviews by Batiza (1996) and Forsyth (1992) and references therein. For an in-depth introduction to MOR processes, see Searle (2013). While much is now known about the structure of backarc spreading centres, we do not consider segmentation in this

setting. We also do not discuss the extensive literature on sub-areal rift systems, but rather provide this review of MOR segmentation as a reference for future comparative studies.

In the next sections, characteristics of tectonic and magmatic segmentation that provide further background framework for this review are described briefly.

Characteristics of tectonic segmentation

The largest-scale physical segmentation of the MOR is defined by the ocean basins, major tectonic plates and regional gradients in ridge properties that extend for thousands of kilometres along the plate boundaries and that are attributed to the influence of deep mantle upwellings or downwellings. Prominent areas of upwelling include hot-spots, such as Iceland and the Azores along the northern MAR (Schilling *et al.* 1983). An example of presumed downwelling, or coldspot, is the Australian–Antarctic Discordance (AAD) along the Southeast Indian Ridge (SEIR) (e.g. Klein *et al.* 1991; West *et al.* 1994; Christie *et al.* 1998).

Superimposed on this regional-scale segmentation is the structural segmentation defined by TFs and NTOs that disrupt the continuity and linearity of the ridge axis (Fig. 1, Table 1). Macdonald *et al.* (1988) proposed a classification scheme for tectonic offsets of orders one to four with large offset TFs classified as first-order discontinuities and the smallest offsets as fourth order. TFs offset the ridge axis by distances of >30 km, partition the MOR at intervals of *c.* 300–500 km, and persist for 10^7 years or more. Second-order discontinuities include large offset OSCs (*c.* 2–30 km) at fast-spreading ridges (e.g. Macdonald *et al.* 1984) and NTOs (with ridge offset of <30 km) of the axial valley at slow-spreading ridges (e.g. Sempéré *et al.* 1990, 1993; Grindlay *et al.* 1991). These features subdivide the ridge at shorter intervals of *c.* 20–100 km at slow-spreading ridges and *c.* 50–250 km at fast-spreading ridges, and persist for *c.* 10^6 years or more. This scale of segmentation corresponds with the ‘spreading cell’ segmentation described above (Schouten & Klitgord 1982; Whitehead *et al.* 1984; Francheteau & Ballard 1983; Crane 1985; Schouten *et al.* 1985). At fast-spreading ridges small OSCs, and the discontinuities identified as devals by Langmuir *et al.* (1986), were classified by Macdonald *et al.* (1988, 1992) as third- and fourth-order discontinuities based on offset length and longevity, with third-order offsets believed to persist for 10^4 – 10^5 years and fourth-order offsets for perhaps a few 1000 years or less. Based on these criteria and the high-resolution bathymetry now available for the EPR, the majority of the offsets identified

Table 1. *Characteristics of Ridge-axis Discontinuities*

Order	Spreading rate	Terminology*	Offset length (km)	Discontinuity morphology	Off-axis morphology	Longevity
1	All	Transform Fault (TF)	$>30^{\dagger}$	Linear valley and transform ridges parallel to spreading direction	Linear valley and transform ridges parallel to spreading direction	Few Ma to 100s Ma; no along-axis migration; TF can evolve into and from non-transform offsets
2	Fast	OSC, Propagating Rift [‡] , (NTO [¶] , Large OSC)	$c. 2-30^{\dagger}, (1-30)^{\S}$	Overlapping ridges enclosing a basin	Zones of discordant topography, can form v-shaped wake	Few Ma, can evolve from TF, often migrate along axis
2	Intermediate	Overlapping Rift Zone, OSC, PR, NTO [¶]	$2-c. 30$	Variable. Overlapping ridges or rifts – may enclose local basin or high. Non-overlapping ridges or rifts separated by zone of rotated seafloor fabric	Zones of discordant topography, can form v-shaped wake if the NTOs migrate	A few Ma to 10 s of Ma, can migrate along axis, disappear or grow into TF
2	Slow	NTO [¶] , oblique zone, en echelon offset, fault-bounded ridge	$c. 2-30^{\dagger}$	Variable. En echelon offset volcanic axes that may overlap. Basin or a ridge separates segments. Shear zones with faults oblique to spreading normal direction. Detachment faults commonly form near NTOs	Zones of discordant topography, can form v-shaped wake if the NTOs migrate	A few Ma to 10 s of Ma, can migrate along axis, disappear or grow into TF
2	Ultra-Slow	Amagmatic or non-volcanic oblique spreading domains	$c. 3-30$ but accommodated in oblique spreading domain	Variable. Deep and wide oblique-trending basins up to 80 km long connecting orthogonal spreading segments with local shallow volcanic regions. Detachment faults at segment ends	Wide zones of oblique-trending basins	10 s (?) of Ma, can migrate along axis, can disappear or grow into TF
3	Fast	3 rd -order discontinuity, small OSC, (deval, SNOO)	$0.5-2^{\dagger}, (0-1)^{\S}$	Variable. Overlapping axial ridges enclosing small basin. Overlapping Axial Summit Trough (AST) within axial high and no enclosed basin. Bathymetric Saddle Points. Change in orientation of axial high and/or AST of $2^{\circ}->5^{\circ}$ with no offset	Small change in orientation of ridge flank fabric, can leave v-shaped trace of offsets in abyssal hill faults	Perhaps 100 ka to Ma, poorly known due to subtle trace and lack of sufficient high resolution ridge flank bathymetry coverage
3-(4?)	Intermediate	Small NTO [¶]	<2	Variable. Change in ridge trend $>5^{\circ}$. Abrupt transition in morphology of axial zone (e.g. from rifted high to shallow rift). Small offset or transition in eruptive fissure zone within rifted axial high. Offset in AVRs within floor of rift. Offset or jog in rift walls.	Some display small offsets in ridge flank faults	Unknown. Some may persist for 100s ka
4	Fast	4 th -order discontinuity, deval, (SNOO)	$<1^{\dagger}, (<0.2)^{\S}$	Variable. En echelon jog in AST. Overlapping AST. Small change in ridge orientation	In some cases small offset or local narrowing of edge of axial high	100s to 10 000 years

*Current common usage with historical terms included in brackets.

[†]from Macdonald *et al.* (1992).[‡]Note that propagating rifts (PRs) at fast-spreading ridges have historically been classified as 1st-order offsets; ridge offset across some PRs is larger than 30 km.[§]From White *et al.* (2000).[¶]Non-transform offset (NTO).

as devals by Langmuir *et al.* (1986) have now been recognized as longer-lived third-order-type discontinuities (White *et al.* 2006).

In early studies, it was proposed that analogous third- and fourth-order discontinuities could also be identified along the slow-spreading MAR, defining fine-scale segmentation of this ridge (Macdonald *et al.* 1988; Grindlay *et al.* 1991; Sempéré *et al.* 1993). A third-order discontinuity was defined as a gap between two chains of volcanoes within the rift valley floor marking the locus of spreading, and fourth-order was defined as a gap in a single chain of volcanoes. However, given the complexity of the axial zone at slow-spreading ridges, including the broad distribution of volcanic features within a wide valley floor, it is now recognized that it is not possible to consistently identify fine-scale segmentation of the axial zone at these ridges (e.g. Smith *et al.* 1995; Table 1).

Characteristics of magmatic segmentation

Magmatic segmentation refers to lengths of ridge fed by a common magma plumbing system. The plumbing system at MORs extends in depth over several tens of kilometres, from the base of the melting regime in the asthenosphere to crystal mush zones, melt lenses and dykes in the crust (e.g. Sinton & Detrick 1992). Given this large range in depth and the evidence that mantle melt is strongly focused toward the ridge axis, MOR plumbing systems are often envisioned as subdivided into discrete melt channels or pathways, branching and coalescing at various depths within the system (e.g. Langmuir *et al.* 1986; Macdonald *et al.* 1988; Grove *et al.* 1993; Kelemen *et al.* 1995; Kelemen & Aharonov 1998).

The boundaries of magmatic segments are identified along MORs from changes in geophysical properties of the axial zone, such as its density structure inferred from gravity anomalies, its crustal thickness and the geometry and distribution of crustal magma chambers. Magmatic segmentation has also been described based on studies of seafloor rock compositions, which reveal lengths of ridge along which erupted magmas share similar petrogenetic histories (e.g. common source compositions, extents and pressures of melting, and/or extents of crystallization). In general, magmatic segmentation based on geochemical variations is better characterized at faster spreading ridges due to denser sample coverage and better age control than at slower-spreading ridges (e.g. Langmuir *et al.* 1986; Sinton *et al.* 1991). However, magmatic segmentation based on geophysical parameters is more readily defined at slow-spreading ridges where large-amplitude variations in the along-axis

bathymetry, gravity and seismic structure are found (e.g. Kuo & Forsyth 1988; Lin *et al.* 1990; Tolstoy *et al.* 1993).

From regional-scale studies of ridge-axis lava compositions, ocean-basin-scale variations are observed that are believed to arise from differences in mantle history and composition and proximity to hotspots (e.g. Schilling *et al.* 1983). Smaller scales of geochemical segmentation have been defined from detailed sampling studies where similar lava compositions are recovered for discrete lengths of the ridge axis (e.g. Langmuir *et al.* 1986; Sinton *et al.* 1991, 2003) and from systematic variations in composition approaching TFs and NTOs (e.g. Christie & Sinton 1981; Bender *et al.* 1984; Thompson *et al.* 1985). It has been proposed that the different scales of magmatic segmentation may reflect a depth hierarchy of magmatic processes with larger-scale segments resulting from variations in source composition, or mantle melting processes, and finer-scale segmentation linked to shallow-level melt segregation processes (e.g. Langmuir *et al.* 1986; Sinton *et al.* 1991). Based on detailed sampling of the southern EPR, Sinton *et al.* (1991) proposed a hierarchy of primary, secondary and tertiary magmatic segments roughly analogous to the orders of tectonic segmentation defined by Macdonald *et al.* (1988). In this and similar studies, geochemical investigations of magmatic segmentation seek to identify lengths of ridge along which sampled lavas share common magmatic histories, offsets across which abrupt changes in composition occur, or systematic along-axis changes in composition approaching an offset.

Synthesis of observations

In the following sections and accompanying figures, we summarize observations of ridge properties along tectonically defined segments, as well as at tectonic discontinuities of different orders (Table 1) for ridges spreading at rates from fast to ultra-slow. Ridge properties discussed include axial morphology, seafloor geology, axial gravity anomalies, seismic velocity structure and crustal thickness, earthquake distributions and lava geochemistry. Each of these properties provides information on different aspects (and time-scales) of the distribution and delivery of melt from the mantle to form the oceanic crust. The morphology of the axial zone (i.e. rift valley or axial high) is closely linked to the thermal structure and strength of the axial lithosphere (e.g. Chen & Morgan 1990). Seafloor depth and gravity anomalies along the ridge axis reflect the integrated density structure of the crust and upper mantle, which are linked to the axial thermal structure, crustal thickness and

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lithology. Geological studies of lava flow types and ages, and faults and fissures along the ridge axis provide constraints on magmatic and tectonic processes and rheological properties. Seismic methods are the primary techniques used to detect the presence of melt and higher temperatures in the crust and mantle and also provide constraints on crustal thickness. Geochemical studies of seafloor samples provide constraints on the composition of the mantle source, and on processes of mantle melting, melt segregation and shallow-level fractionation. While some of these properties (axial morphology, gravity, crustal thickness) reflect time-averaged magma supply to the ridge on 10^3 – 10^6 year time-scales (variable depending on spreading rate), others reflect magmatic processes from present day to *c.* 100-year time-scales (seismic observations of melt in the crust, geochemical studies of young lavas, seismicity studies).

With this review, we focus our discussion on a number of long-standing questions including:

- What are the geophysical and geochemical signals associated with ridge axis discontinuities and, in particular, the smallest tectonic discontinuities that are better located in modern seafloor mapping data, and do these signals scale with size of ridge offset?
- Do tectonic segments defined by these discontinuities coincide with distinct units in other ridge properties indicative of the distribution of magma in the subsurface?
- Are there systematic gradients in ridge properties within tectonic segments consistent with models of central magma supply (and, if so, over what length scales)?
- Do existing observations support predictions of hierarchical models that longer tectonic segments exhibit larger amplitude variations in ridge properties that result from deeper melt generation or focusing processes?

Fast- and superfast-spreading ridges

Fast-spreading MORs (80 – 150 mm a^{-1} , DeMets *et al.* 1994) include the northern and southern EPR and portions of the Pacific–Antarctic Ridge. The best-studied regions to date are the northern EPR from 8 – $18^\circ N$ and the southern EPR from 13 – $21^\circ S$, and we focus our discussion on these areas (Figs 2–4). Along fast-spreading ridges, the region where new crust forms is marked by an elongate axial high, typically 8 – 10 km wide, that rises 200 – 400 m above surrounding seafloor (e.g. Lonsdale 1983; Macdonald *et al.* 1984, 1992; Figs 2 & 3). Active volcanism and hydrothermal venting are narrowly focused along the crest of this axial high, usually within a shallow depression (a few tens of

metres deep) known as the axial summit trough (Fig. 3; Macdonald & Fox 1988; Haymon *et al.* 1991; Fornari *et al.* 1998).

Geophysical properties of discontinuities and ridge segments

Transform faults and first-order segments. The TFs that offset the EPR range from 50 km to 450 km in length and segment the axis into first-order segments *c.* 120 – 600 km long, with several stretches over 1000 km long (Table 1). Approaching a TF, the ridge axis typically deepens by a few hundred metres over a distance of 5 – 10 km (Fig. 2g) and the cross-sectional area of the axis, a measure of the width of the axial high, decreases (Fig. 2f; Scheirer & Macdonald 1993). In some areas, Mantle Bouguer anomalies (MBA), which are gravity anomalies calculated to remove the effects of seafloor topography and a constant thickness and density crust, are somewhat higher (5 – 10 mGal) approaching TFs, suggesting thinner crust and/or cooler/denser crust and mantle (e.g. Orozco, and Clipperton TF (Fig. 2f, Canales *et al.* 2003; Cormier *et al.* 2011) and Garrett TF (Fig. 4f, Magde & Detrick 1995)). Gregg *et al.* (2007), however, report lower residual MBA values within some EPR TF domains including the Siqueiros TF, consistent with thicker, not thinner, crust. A seismic tomographic study spanning *c.* 40 km to either side of the Clipperton TF suggests differences in crustal thickness of no more than 300 m across this transform (van Avendonk *et al.* 1998). However, a more regional seismic tomography study shows that the crust gradually thins by up to 1 km over *c.* 40 km towards both the Clipperton and Siqueiros TFs on the east flank of the ridge (Canales *et al.* 2003).

Across all EPR TFs, a step in the regional depth of the ridge axis is observed, suggesting differences in magma supply to the two adjacent segments. For example, average axial depths along segments bounding the Siqueiros, Clipperton and Orozco TFs differ by several hundreds of metres and differences in the thickness of the crust (of 1 – 2 km) and/or density of the crust and mantle are inferred, assuming isostatic support for this topography (Figs 2 & 3, Macdonald *et al.* 1992; Weiland & Macdonald 1996). Carbotte *et al.* (2004) noted that the shallower segment across all EPR TFs corresponds with the ‘leading segment’, the segment displaced in the migration direction of the ridge axis relative to the underlying asthenosphere, where an asymmetry in melt production may lead to enhanced melt supply (Katz *et al.* 2004).

Within first-order ridge segments, the long-wavelength deepening of the seafloor toward TFs as depicted in early models (Fig. 1) is not observed at many TFs. Instead, seafloor depths along the axis

Fig. 2. Variations in ridge properties along the northern East Pacific Rise (EPR). **(a)** Seafloor bathymetry for the EPR from 8° to 18° from the GMRT synthesis (Ryan *et al.* 2009). All first- and second-order discontinuities are labelled (OSC, overlapping spreading centre; NTO, non-transform offset; TF, transform fault). In lower panels, shaded vertical bars show the locations of transform faults. Bold vertical lines mark the locations of second-order discontinuities, lighter dashed lines mark third-order discontinuities discussed in the text. **(b, c)** $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and K/Ti values for ridge axis basalts providing measures of source enrichment. **(d)** Na_8 values, reflecting extent of mantle melting (Klein & Langmuir, 1987). **(e)** MgO (wt%) reflecting extent of crystal fractionation and temperature of erupted lavas. Geochemical data in panels (b–e) are from C. Langmuir (pers. comm.) and the PetDB synthesis database

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may remain nearly constant to within a few kilometres of the bounding TFs (e.g. south of the Clipperton (Figs 2 & 3) and Garrett TFs (Fig. 4)). In some cases, including south of Clipperton, the ridge axis appears to 'overshoot' the transform domain with volcanic ridges extending on to older crust across the TF (e.g. Fig. 3a; Barth *et al.* 1994).

Prominent examples where long-wavelength gradients are observed include north of Clipperton and south of Orozco where the EPR axis gradually deepens and narrows over many tens of kilometres approaching these TFs (Fig. 2f, g); these changes in ridge morphology have been attributed to a gradual reduction in magmatic budget in these regions (e.g. Macdonald *et al.* 1992). North of Clipperton, evidence supporting this hypothesis includes the finding that no axial magma lens (AML) has been detected within c. 70 km north of the TF, although seismic reflection studies image an AML 1–2 km beneath the seafloor along much of the EPR between 8°30'N and 17°N (Fig. 2g, Detrick *et al.* 1987, 1993; Kent *et al.* 1993, 2000; Babcock *et al.* 1998; Carbotte *et al.* 2000, 2013). Further, lavas erupted over the past c. 1 Ma at the segment north of Clipperton have lower Mg#s than those from the south (Mg# of 5–7 compared with 7–9), indicating lower eruption temperatures and less melt replenishment or greater cooling for the northern segment (Fig. 2e; Batiza *et al.* 1996). Last, the abyssal hills on the ridge flanks are taller (70–80 m) and narrower (1.4–2 km) north of Clipperton than those formed at the shallow broad EPR to its south (30–50 m high, and 2–2.5 km wide, Goff 1991; Fig. 3a, b) indicating long-lived differences in crustal accretion processes that contribute to the formation of abyssal hills. Yet despite these various observations suggesting reduced magma supply north of Clipperton, available seismic data indicate little difference in crustal thickness across the transform (van Avendonk *et al.* 1998). It may be that the differences in ridge morphology across this TF are linked more to differences in crustal density associated with iron enrichment rather than crustal thickness, as has been inferred near the 9°N OSC (Toomey & Hooft 2008).

Second-order discontinuities and segments. Second-order segments along the EPR range from c. 50 to 230 km with an average length of 140 km (Macdonald *et al.* 1988, 1992). In early studies, described previously, it was proposed that these segments are underlain by focused centres of melt upwelling with lateral redistribution of melts towards the bounding discontinuities, perhaps contributing to the along-axis migration of OSCs. Studies conducted over the past 30 years show that while second-order discontinuities are sites of local disruption in the distribution and supply of melt to the ridge, regional along-axis gradients in ridge properties within second-order segments consistent with a gradual reduction in magma supply towards the bounding offsets are not typically observed (Figs 2, 4). The ridge axis usually deepens and narrows within a zone c. 10 km from a second-order discontinuity, and MBA values may be slightly higher (few mGal), consistent with locally reduced magma supply (e.g. at the 9°N and 15°55'S OSCs; Figs 2 & 4; Scheirer & Macdonald 1993). Distinct volcanic mound fields cover the seafloor at OSCs that are thought to result from eruptions at low effusion rates and have been attributed to the presence of small short-lived magma sources that may be widely distributed across the discontinuity zone (White *et al.* 2009). Seismic studies do not show gradual crustal thinning toward second-order OSCs, expected if these were sites of reduced magma supply at the distal ends of regional mantle upwelling centres. Instead, narrow localized zones of thinner crust (by 1–1.5 km) are found within the ridge flank trace of the 9°N OSC, and there is no evidence for thin crust at the other small OSCs that have been imaged (12°54'N, 15°55'S and near 17°S) (Barth & Mutter 1996; Bazin *et al.* 1998; Canales *et al.* 1998).

Interestingly, there is evidence that more melt may accumulate and thicker crust may form near some OSCs for reasons that are not well understood. For example, a wide melt lens (up to 4 km) is found directly north of the 9°N OSC extending to c. 9°17'N, as well as a band of thicker crust (by c. 1 km) centred at 9°15'N (Barth & Mutter 1996; Kent *et al.* 1993, 2000; Canales *et al.* 2003;

Fig. 2. (Continued) (see www.petdb.org for original references), re-normalized for differences in reporting iron content and edited to include samples <2 km from the ridge axis. (f) Mantle Bouguer Anomaly (black line) measured along axis for 8°30'–10°N from J. P. Canales (http://www.marine-geo.org/tools/search/Files.php?data_set_uid=333), and Residual Mantle Bouguer Anomaly (blue line) calculated to remove the thermal effects of plate cooling with age for 12°–16°N from Cormier *et al.* (2011), and for 15°30'–17°N from Weiland & Macdonald (1996). Residual gravity anomaly calculated using seismically inferred crustal thickness is also shown for the 8°30'–10°N region (green line, from J. P. Canales, http://www.marine-geo.org/tools/search/Files.php?data_set_uid=334). Red line shows ridge cross-sectional area from Scheirer & Macdonald (1993). (g) Cross-sections along the axis showing seafloor depth from multibeam bathymetry data and the base of layer 2A, and the axial magma lens (AML) imaged in multi-channel seismic data. AML and layer 2A observations are from Carbotte *et al.* (2013) for 8°30'–10°N, Detrick *et al.* (1987) for 10°30'–12°N, Babcock *et al.* (1998) for 12°30'–13°30'N and Carbotte *et al.* (2000) for 15°30'–17°N. Short vertical arrows mark locations of third-order discontinuities modified from White *et al.* (2006) and Macdonald *et al.* (1992).

Fig. 3. Bathymetric maps illustrating variations in axial morphology and fine-scale segmentation along the northern EPR (full spreading rate 110 mm a^{-1} ; Carbotte & Macdonald 1992). (a) Panels show the EPR north (top) and south (bottom) of the Clipperton Transform Fault. The marked differences in seafloor morphology at these two adjacent segments have been attributed to differences in magma supply to each, with more restricted magma supply to the segment to the north. Thin white line shows location of ridge axis. Black lines show location of cross-axis profiles from (b). (b) The EPR north of Clipperton (profile 1) is associated with a narrow ($<2 \text{ km}$) and deep axial high bordered by high-relief (100–200 m) abyssal hills. In contrast, along the EPR south of Clipperton (profile 2), a broad axial high $\sim 20 \text{ km}$ wide is observed and smooth ridge flank topography with abyssal hills only a few tens of metres high. (c) Close up of the EPR south of Clipperton showing fine-scale segmentation of the axial zone. Third- and fourth-order discontinuities (labelled) are identified from changes in strike of the axial high (third-order) and offsets in the axial summit trough (third- and fourth-order, blue line, Soule *et al.* 2009), which marks the locus of most recent eruptive fissuring and hydrothermal venting. Bathymetric data are from the GMRT synthesis (Ryan *et al.* 2009).

Singh *et al.* 2006a; Combi *et al.* 2008). Seafloor compliance and seismic tomography studies indicate melt volumes in the crust and uppermost mantle under the ridge axis do not diminish toward the 9°N OSC (Crawford & Webb 2002; Toomey *et al.* 2007). Indeed, a local region of higher, not lower, melt content is inferred in the shallow mantle beneath the 9°N OSC (Dunn *et al.* 2001;

Toomey *et al.* 2007). The observations of thicker crust and abundant melt north of the 9°N OSC are intriguing and may provide clues regarding processes driving the long-term southward propagation of this offset (e.g. Canales *et al.* 2003; Klein *et al.* 2013).

There are, however, some regions where long-wavelength changes in ridge properties consistent

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with a gradual reduction in magma supply towards a second-order discontinuity are observed. Along the southern EPR near the 20°40'S OSC, the ridge axis deepens by c. 300 m and the cross-sectional area of the axis decreases over 100 km approaching this offset from the north (Fig. 4). From the gradually increasing axial MBA values, Cormier *et al.* (1995) inferred that thinner crust (by c. 500 m) has been accreted north of the 20°40'S OSC for the past c. 1.5 Ma or more (Fig. 4f). Notably, these changes in ridge properties are similar in magnitude and length-scale to those observed north of the Clipperton TF (cf Figs 2 & 4).

There is also evidence in some locations for significant melt focusing beneath a second-order segment. The shallow broad segment between the Orozco TF and the 16°20'N OSC is associated with a pronounced MBA low (Fig. 2f), indicating locally thicker crust (by as much as 2.25 km) and/or reduced crust/mantle densities beneath the central part of this segment (Weiland & Macdonald 1996). However, a small seamount chain extends west of this segment and enhanced magma supply in this region is presumably linked to the interaction of the ridge axis melt regime with that for the ridge flank seamount chain (e.g. Carbotte *et al.* 2000; Donnelly 2002). Some studies suggest the presence of a melt upwelling centre at c. 9°50'N, with along-axis melt flow away from this region (Barth & Mutter 1996; Wang *et al.* 1996), although other interpretations have also been advanced (Toomey & Hooft 2008).

As observed across TFs, adjacent second-order ridge segments often differ significantly in average depth and cross-sectional area, and the differences in these characteristics can be as large as those observed across TFs (e.g. 11°45'N and 16°20'N; Fig. 2; Scheirer & Macdonald 1993). At most OSCs, the leading segment is the shallower and broader segment, as also observed at EPR TFs (Carbotte *et al.* 2004).

Third-order discontinuities and segments. The third-order segmentation of the EPR is defined by discontinuities that offset the ridge axis by less than c. 2 km (Table 1). These discontinuities take a variety of forms including small OSCs, steps in the axial summit trough with lateral offsets of 0.5–2 km, or small bends in the orientation of the axis (Figs 2 & 3; Langmuir *et al.* 1986; Macdonald *et al.* 1988, 1992; White *et al.* 2006). These discontinuities may persist for tens to hundreds of thousand of years or more, as evident in subtle disruptions in the continuity of seafloor abyssal hills on the ridge flanks (Smith *et al.* 2001; White *et al.* 2002, 2006) and subdivide the ridge into segments c. 20–80 km long (30 km on average). The ridge axis locally deepens (c. 30–100 m) at third-order

discontinuities, and small inflections in the ridge cross-sectional area profile are observed that are smaller than those observed at second-order offsets (Figs 2 & 4; Scheirer & Macdonald 1993). Third-order offsets are not associated with increases in MBA values, suggesting no change in crustal thickness and/or density (Figs 2 & 4).

Within third-order segments, studies from both the northern and southern EPR show that low effusion-rate pillow lavas and lava domes predominate near the bounding discontinuities, and higher effusion-rate sheet flows are found within segment centres (White *et al.* 2000, 2002). Systematic variations are observed in other seafloor parameters including the density of hydrothermal vents and biological communities, suggesting magmatic heat sources are focused near the centre of third-order segments (Haymon & White 2004). White *et al.* (2000, 2002) concluded that each third-order segment has a common crustal-level volcanic plumbing system, analogous to a 'volcanic system' in Iceland, consisting of a central volcano and associated dykes and fissures.

Recent seismic reflection observations of the EPR between the Siqueiros and Clipperton TFs reveal disruptions in the crustal magma lens coincident with third-, as well as the finer-scale fourth-order seafloor discontinuities (Carbotte *et al.* 2013; Marjanovic 2013; Han *et al.* 2014). The nature of magma lens disruptions varies, but in some locations offset and overlapping melt lenses are imaged that in plan view display a geometry similar to that of OSCs (at 9°36'–39'N, Han *et al.* 2014). In addition, third-order discontinuities often coincide with abrupt changes in the depth, width and presence of the AML, whereas this melt body varies over a narrower range in depth and width within individual third-order segments (Kent *et al.* 1993; Babcock *et al.* 1998; Carbotte *et al.* 2000; Marjanovic 2013, Fig. 2g). These observations suggest a partitioning of the crustal magmatic system on these scales, consistent with the hypothesis of White *et al.* (2000, 2002) that third-order segments are the primary volcanic segments of the EPR.

A 3D seismic tomography study of uppermost mantle structure between the Clipperton and Siqueiros TFs reveals pinches and swells in the plan view distribution of low seismic velocities attributed to pooled mantle melts within the shallowest mantle (Toomey *et al.* 2007). Local zones of lower seismic velocity (by 3–4%) attributed to higher melt contents are spaced roughly coincident with the third-order segmentation of the ridge. These tomography results indicate that discrete mantle melt reservoirs underlie third-order segments along the EPR. In addition, reflection images of seismic Moho from the EPR at 9°42'N–9°57'N indicate changes in the nature of reflection Moho

Fig. 4. Variations in ridge properties along the fast-spreading southern EPR. **(a)** Seafloor bathymetry for the EPR from 13° to 21°30'S from the GMRT synthesis (Ryan *et al.* 2009). First- and second-order discontinuities are labelled. In panels below, shaded vertical bar show the locations of transform faults. Bold vertical lines mark the locations of second-order discontinuities and light dashed lines mark locations of geochemically significant third- and fourth-order discontinuities (from Sinton *et al.* 1991). **(b–e)** Geochemical data, renormalized and culled as described in the caption to Figure 2; data sources are from the PetDB synthesis database (primarily from Mahoney *et al.* (1994) and

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from one segment to the next (Aghaei *et al.* 2014) and further support a subcrustal source for this segmentation. The general correspondence of third-order tectonic segmentation with partitioning of the zone of mantle melt accumulation below the crust is inconsistent with early models that called upon large-scale redistribution of magma from regional upwelling centres beneath second-order segments of the EPR (Fig. 1). Instead the observations support more closely spaced sources of mantle melt that roughly coincide with the third-order tectonic segments.

Fourth-order discontinuities and segments. Fourth-order discontinuities identified along the EPR correspond with small steps (<0.5 km) or changes in the trend of the narrow axial summit trough or ridges of pillow lava that mark the locus of active fissure eruptions along the ridge axis (Table 1, Fig. 3c). Due to the subtle seafloor expression of these offsets and limitations in coverage and navigation of older seafloor mapping data, interpretation of these smallest discontinuities has evolved as mapping data have improved (e.g. Haymon *et al.* 1991; Macdonald *et al.* 1992; White *et al.* 2002, 2006). These discontinuities are now best identified along the northern EPR from 8°24'N to 11°45'N (White *et al.* 2006; Soule *et al.* 2009) where they are associated with small (<30 m) local deep points along the axis. Fourth-order offsets are not associated with anomalies in the axial MBA or with changes in the cross-sectional area of the ridge as measured in regional-scale bathymetry studies (Scheirer & Macdonald 1993). However, they often coincide with breaks and local deep points (on the order of 50 m) in the AML and are attributed to fine-scale (5–15 km) segmentation of this melt reservoir (Carbotte *et al.* 2000, 2013; Marjanovic 2013).

Fourth-order discontinuities have also been found to be boundaries separating segments with distinct properties of the eruptive fissure zone. For example, the depth and width of the narrow axial summit trough often change at fourth-order offsets (Macdonald & Fox 1988; Haymon *et al.* 1991). Variations in the dimensions of this structure are interpreted to reflect changes in the history of eruptions and lava effusion rate from one fourth-order segment to the next, with a wider axial summit trough in regions of less frequent eruptions where

ongoing dyking without eruption widens this depression (Fornari *et al.* 1998; Soule *et al.* 2009). These observations support the notion that fourth-order segments are fed by discrete magma reservoirs within the crust each with its own history of eruption and replenishment.

Variations in crustal and uppermost mantle structure associated with fine-scale segmentation was investigated in an early seismic tomography study focused on a small segment bounded by the third-order discontinuity at the EPR c. 9°36'N and what was thought at the time to be a smaller offset at 9°28'N (Toomey *et al.* 1990; Dunn *et al.* 2000). This study found that the zone of low seismic velocities in the mid-lower crust, attributed to higher melt fractions and warmer temperatures, narrows beneath both of these regions. The findings of this study have been used to argue that the finest (fourth-order) scales of seafloor segmentation are related to closely spaced sites of melt injection from the mantle. It is important to note, however, that the results of this tomography study are equivocal. The discontinuity interpreted at 9°28'N is not evident in modern GPS navigation bathymetry data from the region (e.g. White *et al.* 2002; 2006) or in the crustal magma lens (Detrick *et al.* 1987; Kent *et al.* 1993; Carbotte *et al.* 2013), and the apparent pinching of the axial low velocity zone evident in the seismic tomography study occurs near the edge of the survey box where resolution is limited (Dunn *et al.* 2000). Furthermore, the geometry of the region of low velocities and presumed high melt content within the uppermost mantle resolved in this study is inconsistent with that found in the recent, more comprehensive, regional study of Toomey *et al.* (2007). The modern tomography study indicates segmentation of mantle melt on the larger third-order scale, not the smaller scales interpreted from this early study.

Geochemical properties at discontinuities and along ridge segments

Relationships between geochemical and tectonic segmentation along fast- and superfast-spreading ridges have been explored in a number of studies (e.g. Bender *et al.* 1984; Thompson *et al.* 1985; Langmuir *et al.* 1986; Sinton *et al.* 1991; Mahoney *et al.* 1994; Perfit *et al.* 1994; Batiza 1996; Perfit & Chadwick 1998; Castillo *et al.* 2000; White *et al.*

Fig. 4. (Continued) Sinton *et al.* (1991); see www.petdb.org for full references). Geochemically defined segments G–N from Sinton *et al.* (1991) are labelled in (b). (f) Mantle Bouguer Anomaly (black line) (Cormier *et al.* 1995; Magde & Detrick 1995). Light line shows ridge cross-sectional area from Scheirer & Macdonald (1993). (g) Cross-sections along the axis showing seafloor depth from multibeam bathymetry data and the base of layer 2A, and the AML imaged in multi-channel seismic data (Hooft *et al.* 1997). Short vertical arrows mark locations of third-order discontinuities from Scheirer *et al.* (1996).

2000, 2002; Smith *et al.* 2001; Sims *et al.* 2002). Below we summarize key geochemical findings along the northern and southern EPR (Figs 2 & 4).

Transform fault-bounded and second-order non-transform segmentation. Accumulating geochemical data suggest that along fast-spreading ridges, TFs and second-order discontinuities (large OSCs) share many of the same chemical systematics (Langmuir *et al.* 1986). We therefore summarize these observations together.

Chemical indicators of extents of melting, such as $\text{Na}_{8,0}$ (wt% Na_2O corrected for fractionation; Klein & Langmuir 1987), along fast-spreading ridges commonly show gradual, modest along-axis undulations over hundreds of kilometres (as do indicators of pressures of melting). Based on present sampling of the northern EPR, average $\text{Na}_{8,0}$ appears to vary relatively smoothly from south of the Siqueiros TF to the $11^\circ 45'\text{N}$ OSC, while crossing both the 9°N OSC and the Clipperton TF (Fig. 2d). While source heterogeneity has been shown to produce local chemical variability along the EPR (e.g. Langmuir *et al.* 1992), the observation of long-wavelength average $\text{Na}_{8,0}$ variations along lengths of the EPR where K/Ti is relatively constant (e.g. north and south of Clipperton TF) suggests that mantle temperature may vary gradually (although modestly) over these spatial scales, leading to gradual variations in average extents of melting along axis. This argues against an earlier suggestion, based on limited sampling adjacent to the Tamayo TF on the northern EPR, that basalts erupted near transforms in general are produced by smaller extents of melting due to a cold edge effect (Bender *et al.* 1984). Superimposed on these long-wavelength variations in extents of melting are local perturbations often associated with offsets. At the $11^\circ 45'\text{N}$ OSC, for example, there is an abrupt change in $\text{Na}_{8,0}$, suggesting that magmas north of the OSC are produced by greater extents of melting (lower $\text{Na}_{8,0}$) than those to the south (Fig. 2d).

With respect to source composition, in places along the northern and southern EPR where abrupt changes in isotopic or incompatible trace element ratios occur, these changes are often associated with TFs or large-offset OSCs. Across the Orozco TF, for example, K/Ti ratios change abruptly at relatively constant MgO wt% (Fig. 2c), indicating derivation from distinct mantle sources across the offset (Donnelly 2002). A similarly abrupt change in Sr isotopic composition occurs across the $20^\circ 40'\text{S}$ OSC (Mahoney *et al.* 1994, Fig. 4b). However, not all TFs or large-offset OSCs coincide with significant changes in source composition (e.g. Clipperton and Garret TF; Figs 2, 4). Overall, the observation of long-wavelength gradients in isotopic composition (Fig. 4b) and both the presence

and absence of abrupt changes in source composition at TFs and second-order discontinuities argues against the early notion that ridge segments bounded by these discontinuities in general represent coherent geochemical units derived from a common source composition (e.g. Langmuir & Bender 1984).

Superimposed on variations in source composition and melting conditions are the often prominent along-strike changes in extent of crystal fractionation. Abrupt changes in MgO wt%, a measure of magma temperature and extent of crystallization, occur across the Clipperton and Orozco TFs, and the 9°N and $20^\circ 40'\text{S}$ OSCs (Figs 2e & 4c). Such changes in extent of fractionation have been attributed to variations in the balance between magma supply and cooling rate (e.g. Christie & Sinton 1981). Little change in fractionation, however, is observed across the Garrett TF or the $16^\circ 20'\text{N}$ OSC. The relatively primitive nature of lavas on either side of the Garret TF argues against the suggestion that magmas erupted near transforms are typically more fractionated (e.g. Bender *et al.* 1984). Interestingly, however, at some second-order offsets, significant extents of crystallization have been observed, most notably at the 9°N OSC, where highly fractionated (low MgO wt%) magmas are sampled (Fig. 2e) (Wanless *et al.* 2010; Klein *et al.* 2013).

Third-order discontinuities and segments. Characterization of geochemical systematics associated with third-order (and fourth-order) discontinuities requires fine-scale sampling and good spatial control. Based on samples recovered at an average spacing of c. 8 km along the EPR from 5.5°N to 14.5°N , Langmuir *et al.* (1986) identified significant chemical variability associated with even small ridge offsets, most of which are now classified as third-order offsets (White *et al.* 2006). Further, they raised the possibility that the ridge segments in between these offsets may exhibit coherent chemical systematics distinct from adjacent segments, although the sampling density in their study was insufficient to fully explore this question.

Sinton *et al.* (1991) identified similar chemical variability associated with small ridge offsets along the southern EPR from $13\text{--}23^\circ\text{S}$ (Fig. 4), and demarcated lengths of ridge (averaging c. 69 km) that appear to have common parental magma compositions likely resulting from similar melting conditions. They described these lengths of ridge as 'secondary magmatic segments' (labelled G-N in Fig. 4b) to distinguish them from their 'primary magmatic segments' based on isotopic composition. The boundaries of these segments correspond primarily with third-order offsets (mostly small OSCs), in addition to the Garret TF and two second-order

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offsets. Sinton *et al.* (1991) suggested further that these findings could be used to place constraints on the along-axis spacing of mantle melt segregation events. They argued that the spacing of injection centres could be less than the lengths of the geochemically defined segments (e.g. if two adjacent injection centres happen to produce similar parental magma compositions), but they cannot be greater.

Few studies have collected lavas at sufficient sampling density to explore geochemical variations at individual third-order offsets and adjacent segments in detail. In a study of the third-order discontinuity at 9°37'N on the EPR (Figs 2 & 3), Smith *et al.* (2001) showed that the southward-propagating east limb of this discontinuity erupts primitive lavas of relatively uniform composition, while the west limb erupts lavas that are both more fractionated and more variable in parental magma composition, suggesting less efficient magma mixing beneath the western dying limb. These findings support the idea that the segments adjacent to these third-order offsets may have isolated magma plumbing systems fed by compositionally variable melts (e.g. Langmuir *et al.* 1986; Reynolds *et al.* 1992; Perfit *et al.* 1994; Smith *et al.* 2001; White *et al.* 2002; Bergmanis *et al.* 2007).

Fourth-order discontinuities and segments. In one of the few studies of sufficient sampling density to explore detailed geochemical systematics associated with fourth-order offsets, Goss *et al.* (2010) report subtle differences in the compositions of lavas sampled along the fourth-order segments north and south of the 9°53'N deval resulting from differences in extents of melting and fractionation. Carbotte *et al.* (2013) showed that the boundaries between these small-scale geochemical and tectonic segments in this region correlate with disruptions in the continuity and characteristics of the underlying melt lens. Based on these relationships, they argued that the fourth-order tectonic segmentation is inherited from partitioning of the crustal magma lens, and that the chemical differences along adjacent segments result from dominantly vertical transport of melt from these chemically distinct melt lens segments.

Intermediate spreading ridges

Intermediate spreading centres spread at rates from *c.* 50 to 80 mm a⁻¹ (DeMets *et al.* 1994) and include the Galapagos Spreading Centre (GSC), the Juan de Fuca (JdF) and Gorda Ridges, the Chile Rise, the SEIR and portions of the Pacific–Antarctic Ridge. Here we focus primarily on observations from the SEIR 100–111°E (Figs 5 & 6) and the western GSC 98–91°W (Fig. 7), two well-studied regions

where the range of typical intermediate spreading ridge structure is found.

While the morphology of intermediate spreading ridges varies widely from the axial highs characteristic of fast-spreading ridges to the >1 km-deep rift valleys typical of slow-spreading ridges (e.g. Small 1998), the most common morphological types are rifted axial highs and shallow rift valleys (Fig. 6, Cochran & Sempéré 1997; Goff *et al.* 1997; Sempéré *et al.* 1997; Small *et al.* 1999; Sinton *et al.* 2003; Christie *et al.* 2005). Rifted axial highs are characterized by a narrow depression 50–200 m deep at the summit of an axial high that is a few kilometres wide (Fig. 6c, segment P3). Shallow valleys are typically 2–4 km wide and <400 m deep (Fig. 6a, segment S1). Abrupt changes in the morphology of the ridge axis from axial highs to shallow rifts are an important characteristic of intermediate spreading ridges and are observed within both the SEIR and GSC study regions. These abrupt transitions are believed to result from a threshold response of the ridge to small changes in magma supply: as magma supply decreases below a certain threshold, the reservoir of magma in the crust rapidly disappears, leading to a large increase in the thickness/strength of the axial lithosphere, which in turn controls the nature of axial topography (Phipps Morgan & Chen 1993; Cochran & Sempéré 1997). Along many intermediate spreading ridges, regional gradients in ridge properties are observed that extend over hundreds to thousands of kilometres and are attributed to gradual variations in magma supply with proximity to local hotspots (e.g. the Galapagos hotspot near the GSC; Schilling *et al.* 1982) or cold-spots such as AAD along the SEIR (Klein *et al.* 1991; West *et al.* 1994; Christie *et al.* 1998). In addition to these regional gradients, 'tectonic corridors' have been identified from systematic differences in the rate and asymmetry of across-axis seafloor subsidence with age that span multiple TF-bounded segments (Hayes & Kane 1994). Along the SEIR, five tectonic corridors with average lengths of 800 km are identified, one of which extends from 100°–115°E and includes the area of Figure 5.

Geophysical properties of discontinuities and ridge segments

Transform faults and first-order segments. The SEIR is offset by numerous long-lived TFs that range in length from 37 to 130 km and subdivide the ridge into segments 55–700 km long (Small 1998; Table 1). The ridge axis typically deepens by several 100 m within a few kilometres of these TFs, and axial MBA values are typically 5–10 mGal higher (Fig. 5g, f; Cochran & Sempéré 1997). Seismic studies indicate the crust is slightly thinner

Fig. 5. Variations in ridge properties along the intermediate spreading South East Indian Ridge (SEIR). **(a)** Seafloor bathymetry of the SEIR from 100°E to 111°E from Baran *et al.* (2005) and the GMRT synthesis (Ryan *et al.* 2009). Second-order segments P1 through S1 are labelled. In panels below, shaded vertical bars show the locations of transform faults. Bold vertical lines mark the locations of OSC discontinuities. **(b–e)** Geochemical data, as in Figures 2 and 4, from PetDB. **(f)** Mantle Bouguer Anomaly measured along axis from Cochran & Sempéré (1997). **(g)** Cross-section

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(<0.5 km) near TFs (Holmes *et al.* 2008). In general, the local anomalies in ridge properties near TFs are similar in magnitude to those observed at the second-order discontinuities along this ridge. The depth of the ridge axis along the SEIR within the region of Figure 5 is dominated by long-wavelength deepening toward the AAD to the east. Superimposed on this regional gradient are steps in the depth of the ridge axis at TFs. As observed at fast-spreading ridges, there is a marked asymmetry in the depth of the two adjoining ridge segments across TFs, with the leading segment consistently shallower than the trailing segment (Carbotte *et al.* 2004).

In marked contrast to the SEIR, few TFs are found along the hotspot-dominated GSC. Indeed, the Galapagos TF located at *c.* 90°45'W is the only transform offset for over 1500 km along the GSC from the western tip of the GSC to the Inca TF at 85°30'W.

Second-order discontinuities and segments. The large non-transform or second-order offsets found along intermediate spreading ridges display a wide range of morphological types (Cochran & Sempéré 1997; Goff *et al.* 1997; Sempéré *et al.* 1997; Small *et al.* 1999; Sinton *et al.* 2003; Christie *et al.* 2005). In some cases, overlapping shallow rifts or axial ridges are observed that curve toward each other enclosing an elliptical overlap basin (e.g. 93°15'W on the GSC; Sinton *et al.* 2003; Fig. 7) or a local topographic high (e.g. the P3-P4 offset at 104°20'E on the SEIR; Baran *et al.* 2005). At other discontinuities the offset ridges or rifts do not overlap but exhibit an oblique jog in the axial zone or a region of locally deeper seafloor cross-cut by faults that are highly oblique to the ridge axis trend (e.g. the PR at 95°30'W on the GSC; Fig. 7). Transitional discontinuities are also observed such as the *c.* 30 km offset at 102°50'E on the SEIR. Here a narrow ridge-perpendicular transform fault zone is present, but on the ridge flanks there is a broad zone of disturbed and oblique-trending topography without a ridge-perpendicular fracture zone (Fig. 5a), indicating the TF has developed only recently.

Along the SEIR, second-order segments are typically 60–80 km long (Sempéré *et al.* 1997; Fig. 5). Many display arch-shaped along-axis depth profiles, with variations of a few hundred metres to a kilometre from segment ends to shallow points, which

are located either within the centres or offset to the western ends of these segments. Gravity variations within these segments mirror the variations observed in depth, with axial MBA lows of *c.* 10–20 mGal centred over the shallowest portions of many, but not all, segments (P1, P2, P3 and R; Fig. 5f; Cochran & Sempéré 1997). These segment-scale MBA variations are attributed to the focusing of mantle melt beneath these shallower regions (Cochran & Sempéré 1997). Multi-channel seismic studies reveal an axial magma body in the crust that is only detected beneath the central shallow portions of these same segments (Fig. 5g; Baran *et al.* 2005), supporting the hypothesis that mantle melt supply is focused beneath these regions. Properties of this magma body, including its average depth and width, as well as the average thickness of the volcanic layer of the crust inferred from observations of seismic layer 2A, vary from one ridge segment to the next (e.g. average AML depth is 1480 m at P1 and 2100 m at P2; average layer 2a thickness is 310 m at P1 and 460 m at P2; Baran *et al.* 2005), consistent with the notion that a separate melt supply system underlies each segment. The seismic data also suggest that second-order segments can be further subdivided. For example, within segment P3 a break in the continuity of the magma body coincides with a small ridge offset at 103°20'E (see below; Fig. 5g) along with a change in average depth of the body from 1900 m to 2320 m and in the thickness of layer 2A (from 370 m to 740 m), suggesting that two separate magmatic systems underlie this second-order segment.

Seismic measurements of crustal thickness are available for segments P1, P2 and S1 of the SEIR (see Fig. 5 for location; Holmes *et al.* 2008). These data indicate slightly thinner crust (<0.5 km) near the discontinuities that bound these segments. However, overall crustal thickness variations found along this intermediate spreading ridge are minor (<1 km). Hence, although there is good evidence for melt focusing beneath these second-order segments, the crust formed is quite uniform in thickness along the length of the segment.

Along the GSC, second-order segments are 100 to >200 km long (Sinton *et al.* 2003; Christie *et al.* 2005), considerably longer than along the SEIR and more similar to the lengths of second-order segments on the EPR. Variations in ridge properties within these second-order segments are also more subdued than along the SEIR and are

Fig. 5. (*Continued*) along the axis showing seafloor depth from multibeam bathymetry data, and the base of layer 2A and the AML imaged in multi-channel seismic data from Baran *et al.* (2005). Horizontal bar at top of panel indicates type of axial morphology: (colours in online version of the figure) red, axial high; green, rifted high; blue, shallow rift valley; black, rift valley. The locations of small-scale (third-order) discontinuities found along the axis are indicated with short vertical lines.

Fig. 6. Bathymetric maps illustrating axial morphology of a magma-rich and magma-poor segment and fine-scale segmentation along the SEIR (full spreading rate 76 mm a^{-1}). **(a)** Panels show the magma-rich P1 (left) and magma-poor S1 (right) segments of the SEIR. Thin black lines show location of ridge axis. Bold black lines show location of cross-axis profiles shown in **(b)**. **(b)** Although spreading rates are approximately constant throughout this region, the P1 segment (profile 1) is associated with a broad shallow axial high, while at the S1 segment (profile 2) a shallow axial valley is present. **(c)** Close up of the P3 segment of the SEIR showing the small $103^{\circ}35'E$ discontinuity that marks the regional transition from predominantly rifted axial high topography to the west to shallow rift valley topography to the east. Bathymetric data are from the GMRT synthesis (Ryan *et al.* 2009).

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dominated by regional gradients associated with the Galapagos hotspot (compare Figs 5 & 7). East of c. 98°W, seafloor gradually shoals (Fig. 7h) and axial MBA values gradually decrease by >60 mGals (Fig. 7f; Canales *et al.* 2002) suggesting thicker crust and/or warmer temperatures toward the Galapagos hotspot. These regional gradients are interrupted at some second-order discontinuities (95°30'W and 96°30'W). Within 5–20 km from these offsets the ridge axis locally deepens by 100–300 m, and local gravity highs of 5–15 mGal are observed. However, no local MBA high is associated with the 93°15'W OSC (Fig. 7f).

From the available seismic data, there is no evidence for locally thinner crust at second-order offsets along the GSC (Fig. 7e; Canales *et al.* 2002). Rather, crustal thickness increases gradually toward the hotspot, consistent with inferences from gravity data. MCS data have been acquired along the ridge axis from c. 95°W to 91°20'W and reveal a crustal magma body beneath most of the ridge axis east of 94°20'W (Fig. 7g; Detrick *et al.* 2002; Blacic *et al.* 2004). Although there is a 15 km-long break in the continuity of the AML at the 93°15'W OSC, this discontinuity is not a significant boundary in magma lens properties as might be expected if regional upwelling centres fed the adjoining second-order segments. Notably, a major change in the depth and width of the AML coincides with a much smaller offset at 92°40'W (see below) and not the second-order 93°15'W discontinuity (Fig. 7g; Blacic *et al.* 2004).

Third-order discontinuities and segments. Along intermediate spreading ridges, a class of small offsets is found that are analogous to the third- and possibly fourth-order offsets recognized along the EPR. At the GSC, these small discontinuities correspond with offsets of typically <2 km or bends in the orientation of the axis of >5° (Fig. 7; Sinton *et al.* 2003; Christie *et al.* 2005; White *et al.* 2008). Similar small-scale offsets are observed along the SEIR (Fig. 6) with interpreted locations indicated in Figure 5. Some of these small discontinuities appear to have an identifiable trace on the near-ridge flanks indicating they have persisted at these locations for 10⁵ years or more (e.g. 103°35'E; Fig. 6), whereas others do not have a trace on the near-ridge flanks and appear to be shorter lived. However, whether there are nested scales of fine-scale segmentation like those on the EPR, with longer-lived third-order segments subdivided into shorter-lived segments, has not been established and these offsets are considered here as one class (Table 1).

At the GSC, these small offsets define morphological segments 30 km to >100 km long that are interpreted as the primary volcanic segments along this ridge (Sinton *et al.* 2003; Christie *et al.* 2005;

White *et al.* 2008). The available seismic data from the western GSC show breaks in continuity and changes in the depth of the AML at several of these small offsets (93°53'W, 92°40'W, 91°33'W; Fig. 7g; Detrick *et al.* 2002; Blacic *et al.* 2004), suggestive of separate magmatic systems beneath the adjacent segments. Furthermore, some of these small offsets also mark major regional boundaries in other ridge properties. For example, the transition in the morphology of the ridge axis from an axial high to a rifted high along the western GSC coincides with a small discontinuity at 92°40'W of only 500 m offset (Fig. 7; Sinton *et al.* 2003; White *et al.* 2008). An abrupt step in both the depth (0.6 s two-way travel time or c. 1.5 km; Fig. 7g) and width (from 0.5–1.5 km east of 92°40'W to 0.7–2.4 km to the west) of the AML (Detrick *et al.* 2002; Blacic *et al.* 2004) also occurs at this discontinuity, along with a transition in the nature of seafloor/volcanic relief (Colman *et al.* 2012). Similarly, along the SEIR a small offset of <200 m at 103°35'E coincides with an abrupt transition in ridge morphology from rifted axial highs to shallow rift valleys (Figs 5 & 6) and a c. 10 mGal step in the regional MBA gradient (Fig. 5g; Cochran & Sempéré 1997).

Geochemical properties of discontinuities and ridge segments

Chemical analyses of lavas recovered from the western GSC display long-wavelength gradients in magma composition that, with geophysical and geological variables, correlate to first-order segments (Fig. 7). On the largest scale, isotopic and incompatible trace element ratios reveal a westerly decrease in source enrichment with distance from the Galapagos hotspot that correlates with decreasing crustal thickness and increasing axial depth, MBA and (where investigated) melt sill depth (Schilling *et al.* 1982; Detrick *et al.* 2002; Sinton *et al.* 2003). This overall geochemical trend is subdivided into three >200 km-long geochemical provinces reflecting changes in source enrichment ('primary magmatic segments' of Sinton *et al.* 1991): an enriched-mid-ocean ridge basalt (E-MORB) province east of 92°40'W; a transitional (T-MORB) province between the 92°40'W third-order offset and the 95°30'W PR tip; and a normal (N-MORB) province west of 95°30'W (Fig. 7b, c, Sinton *et al.* 2003; Cushman *et al.* 2004; Ingle *et al.* 2010). These geochemical provinces also coincide with changes in axial morphology (eastern axial high, central transitional and western axial deep morphologies), and the 92°40'W third-order offset coincides with a rapid westward doubling of the depth to the melt lens and the base of layer 2A (Detrick *et al.* 2002). Notably, although the 95°30'W PR forms a boundary in terms of source composition, the

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other second-order offset in this region at 93°15'W (variously described as a PR or OSC) does not. In this sense, the GSC appears to share some similarities with the EPR in that some, but not all, second- and third-order boundaries coincide with abrupt changes in source composition, and magma compositions along some of these segments share common source characteristics.

The strong influence of the Galapagos hotspot on the GSC lavas to 95°30'W makes it difficult to isolate chemical changes resulting from variations in extent and pressures of melting from those resulting from variations in source composition, and thus to explore whether 'secondary magmatic segments' like those described by Sinton *et al.* (1991) for the southern EPR can be identified along the GSC. Cushman *et al.* (2004) suggested that E-MORB and N-MORB may be produced by similar mean extents of melting but that greater amounts of melt are produced near the hotspot due to hydrous melting of a greater mantle volume. They did not, however, explore the relationship between segmentation and along-axis changes in melting conditions.

The most prominent geochemical systematics associated with offsets along the GSC are variations in the extent of fractionation. Sinton *et al.* (2003) identified eight segments, 30–120 km in length, that form an along-axis saw-tooth pattern in Mg#, with relatively smooth variations within segments and abrupt changes across second- and some third-order offsets (Fig. 7d). The boundaries of these fractionation-defined segments also coincide with changes in the characteristics of the melt lens, and the variability of Mg# within segments appears to correspond to variability in melt lens depth and continuity. Sinton *et al.* (2003) attributed the segmentation in fractionation to along-strike changes in the thermal structure of the crust. Noting the similarity of these fractionation trends to the 200 km-long fractionation gradient approaching the 20°40'S OSC on the EPR, they speculate that the segmentation in

Mg# along the GSC may reflect gradients in magma supply, with greatest input of magma at the eastern end of most segments (nearest the hotspot).

Despite the collection of significant geophysical data along the SEIR between 100°E and 111°E (Baran *et al.* 2005; Fig. 6), sampling remains sparse, allowing only general observations to be made. On a regional scale, the SEIR shows long-wavelength gradients in isotopic ratios and other chemical parameters suggesting an eastward decrease in average extents of melting and diminishing multi-source component source enrichment (Mahoney *et al.* 2002; Graham *et al.* 2006; Russo *et al.* 2009; Hanan *et al.* 2013). Within the region shown in Figure 6a, transition occurs near the P3–P4 offset with lower average Na_{8,0} values (indicating greater extents of melting) along segments west of the offset, where the ridge is characterized by an axial high and rifted high morphology and lower MBA values (Fig. 6). In contrast, average Na_{8,0} values in segments east of this boundary are higher, suggesting smaller extents of melting associated with the areas of deeper axial rift valleys and higher MBA. The limited chemical data available thus provide general support for previous work suggesting an abrupt transition in ridge properties in this region (Baran *et al.* 2005), and is consistent with the idea that intermediate spreading ridges display a threshold response to small changes in melt supply (Phipps Morgan & Chen 1993; Cochran & Sempéré 1997).

Slow-spreading ridges

Slow-spreading ridges (20–50 mm a⁻¹) include the MAR, Reykjanes Ridge, Central Indian Ridge and the southern Gorda Ridge. The best-studied slow-spreading ridge to date is the northern MAR, and the following discussion is focused mainly on the region between 15°N and 40°N, which spans five

Fig. 7. Variations in ridge properties along the intermediate spreading Galapagos Spreading Centre (GSC) from 98°W to 91°W. (a) Seafloor bathymetry of the GSC with ridge axis indicated by black dashed line from Canales *et al.* (2003). In panels below, black vertical lines mark the locations of second-order discontinuities including overlapping spreading centres and propagators. Black dashed lines show location of smaller discontinuities identified by Sinton *et al.* (2003) that mark significant boundaries in other ridge properties. (b) ⁸⁷Sr/⁸⁶Sr from axial basalts from Ingle *et al.* (2010). (c) K/Ti values from Cushman *et al.* (2004). (d) Variation in glass Mg# with regions of smoothly varying Mg# shaded. Samples are keyed to chemical affinity based on K/Ti, as follows: N-MORB (blue squares), T-MORB (green circles) and E-MORB (inverted red triangles). Figure from Detrick *et al.* (2002). (e) Crustal thickness estimated from seismic refraction lines (squares with uncertainty bounds) and MCS data (dots) from Canales *et al.* (2003). (f) Along-axis MBA and residual anomaly calculated for crustal thickness from (e), also from Canales *et al.* (2003). (g) TWTT to the AML reflector detected by MCS survey along axis from Blacic *et al.* (2004) and Detrick *et al.* (2002); survey was run off axis between 94°42'W and 94°21'W. (h) Bathymetry, segmentation and axial morphology from Sinton *et al.* (2003). Axial depth profile shows water depth along the present spreading axis. Eastern, Middle and Western indicate three regional provinces identified by Sinton *et al.* (2003) based on axial morphology. Smaller-scale segmentation and axial morphology are shown by the coloured rectangles where red denotes axial high, green denotes transitional and blue denotes axial deep. Sense of offset (not to scale) between the smaller segments is indicated.

Fig. 8. Variations in ridge properties along the northern Mid-Atlantic Ridge (MAR). (a) Seafloor bathymetry for the MAR (15–40°N) from the GMRT synthesis (Ryan *et al.* 2009). Black line marks the ridge axis and its offsets. Locations of detailed bathymetry maps in Figure 9 are shown by white rectangles. The location of the study of Gale *et al.* (2011) is also marked (Fig. 10). Transform faults are labelled. In the panels below, shaded vertical bars show the locations of transform faults. (b–e) Geochemical data, as in Figures 2 and 4, from PetDB. (f) Mantle Bouguer Anomaly (MBA) from Thibaud *et al.* (1998). (g) Along-axis depth profile obtained from multibeam bathymetry data.

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Fig. 9. Detailed bathymetry from the northern MAR illustrating contrasting ridge morphology (full spreading rate 22 mm a^{-1}). (a) Bathymetry for three segments south of the Oceanographer fracture zone. White lines run along the axis of the segments. Black lines show the location of the profiles in (c). Examples of the ‘inside corner’ at a transform fault and NTO are indicated. (b) Bathymetry for the segment north of the Atlantis TF. Continued extension on a long-lived detachment fault has produced the domed and corrugated Atlantis Massif (AM), which sits at the inside corner of the transform–ridge axis intersection. White lines mark the axis, and the black line the location of the profile in (c). (c) Depth profiles across the four segments shown in (a) and (b). Note the shallow depths in OH1, which is considered magmatically robust, and the deeper bathymetry in the AM segment, which is considered magmatically starved. (d) Along-axis profiles for the segments south of the Oceanographer FZ showing deeper segment ends and shallower segment centre. Along-axis profile for the AM segment is not typical. It shallows toward the north with no midpoint high.

TFs (Fig. 8). Figure 9 shows detailed bathymetry for a few ridge segments within this section. Slow-spreading ridges such as the northern MAR are marked by a 1.5–3 km-deep, 30–45 km-wide axial rift valley (Tapponnier & Francheteau 1978; Macdonald 1986). An inner valley floor (5–12 km wide) is bordered by valley walls in which large (hundreds of metres) inward-facing normal faults displace the crust upwards to form the crestal mountains. The inner valley floor, where new ocean crust is accreted, commonly contains an axial volcanic

ridge (AVR) that is considered to be the predominant site of volcanic activity (Ballard & van Andel 1977; Smith & Cann 1993). AVRs can range up to several hundreds of metres high, several kilometres wide and tens of kilometres long.

Geophysical properties of discontinuities and ridge segments

Transform fault (TF) and first-order segments. TFs defining the first-order segmentation of the

MAR offset the ridge axis over varying lengths, with the greatest being the Romanche TF near the equator, which offsets the ridge axis by *c.* 900 km (Bonatti *et al.* 1994; Table 1). As the ridge axis approaches a TF, it deepens by 500–2000 m and MBA values increase, indicating thinner and/or cooler crust and mantle (Fig. 8; Sempéré *et al.* 1993; Detrick *et al.* 1995; Thibaud *et al.* 1998). Approaching some TFs, the AVR breaks down into groups or chains of isolated seamounts suggesting more limited magma supply near the ridge–transform intersection. In other places, however, a robust AVR reaches the transform discontinuity and fills the nodal basin at the intersection (e.g. Smith & Cann 1999).

The inside of the *c.* 90° bend of the plate boundary at both ridge–transform intersections and NTOs is referred to as the inside corner (Fig. 9). There are several differences in the characteristics of inside corners at first-order ridge–transform intersections compared to the opposite outside corners. Commonly, the inside corner flanks have shallower seafloor, higher gravity values, exposures of lower crustal and upper mantle rocks, and no volcanic features (e.g. Severinghaus & Macdonald 1988; Cannat 1993; Tucholke & Lin 1994; Escartín & Lin 1995). These observations are consistent with tectonic thinning of the crust at the inside corner by long-lived (million years or more) detachment faults that can reach horizontal offsets of tens of kilometres and may account locally for 60–100% of plate separation (e.g. Tucholke *et al.* 1998; Ohara *et al.* 2001; Searle *et al.* 2003; Okino *et al.* 2004; Buck 2005; Cannat *et al.* 2006; Smith *et al.* 2006; Baines *et al.* 2008; Blackman *et al.* 2008; Grimes *et al.* 2008; Smith *et al.* 2008). In the detachment fault model, lower crust and upper mantle rocks are exhumed to the seafloor at the inside corner and the volcanic section is spread to the outside corner (Dick *et al.* 1981; Karson 1990; Tucholke & Lin 1994). The formation of detachment faults is likely dependent on a balance between several factors, but magma input is considered a key variable. Numerical modelling suggests that long-lived detachment faults may form primarily when the fraction of plate separation taken up by magma accretion is between *c.* 0.3 and 0.5 (Buck *et al.* 2005; Tucholke *et al.* 2008; Olive *et al.* 2010), and this is supported by their common occurrence near ridge–transform intersections, where melt supply is thought to be reduced (Tucholke *et al.* 2008).

The along-axis topographic profile (Fig. 8g) shows that between TFs there is no regional arch as depicted in the schematic of hierarchical ridge segmentation of (Fig. 1; Macdonald *et al.* 1988). Larger regional-scale gradients are observed, however, including the gradual shoaling of the MAR

as it approaches the Azores and Iceland hotspots. In addition, tectonic corridors have been identified along the southern MAR with similar geophysical and geochemical characteristics that appear to span broader sections of the ridge axis (hundreds of km) and several TFs (Kane & Hayes 1992). There are some transforms, however, that show strong variations across them, such as the Romanche TF, which marks a regional boundary in the geochemistry (Kashintsev *et al.* 2008) and possibly in the thermal structure of the underlying mantle (Bonatti *et al.* 2001).

Second-order discontinuities and segments.

Between TFs along slow-spreading ridges, NTOs divide the axis into second-order segments with lengths of 20–100 km (Fig. 9a, b; Table 1; Sempéré *et al.* 1993; Gente *et al.* 1995; Thibaud *et al.* 1998). NTOs have typical offsets of 15–30 km (e.g. Sempéré *et al.* 1993) and many leave off-axis traces that indicate that some of these offsets migrate up and down the ridge axis resulting in segments with variable lengths (e.g. Gente *et al.* 1995). NTOs at slow-spreading ridges also have diverse morphologies. One type is characterized by a steep-sided ridge (sometimes referred to as a septal ridge) that appears to be fault-bounded (Sempéré *et al.* 1993) and separates the two segments (Sempéré *et al.* 1993; Spencer *et al.* 1997). A second type of NTO is an oblique shear zone, first identified by Sempéré *et al.* (1993) at 24°51'N on the MAR and described as a failed TF. The offset is accomplished by short, *en echelon* zones in which faulting is highly oblique to the normal spreading direction (e.g. Gracia *et al.* 2000). A third type of NTO is marked by a step or offset in the volcanic axis between adjacent segments.

Along the axis of second-order segments, the shallowest depths are typically located near the segment centre, 400–700 m higher than segment ends (Fig. 9d) (e.g. Sempéré *et al.* 1993; Gente *et al.* 1995; Thibaud *et al.* 1998). Thibaud *et al.* (1998) categorized each second-order segment between 15°N and 35°N as 'hot' or 'cold'. Their hot segments, such as the three segments south of Oceanographer TF (e.g. Figs 8 & 9a), show a strong change in along-axis MBA (25–35 mGal), bathymetry (1–2 km) and have a narrow axial rift (*c.* 5–10 km). In contrast, cold segments, such as the one north of Atlantis transform (e.g. Figs 8 & 9b), have a smaller change in along-axis MBA (3–17 mGal), relief (hundreds of metres) and a wide axial rift (*c.* 15–20 km). The shallow axial depths near segment centres, as argued by Neumann & Forsyth (1993), likely represent a combination of hot and thick crust and thinner lithosphere than at deeper segments, and these shallow segments are commonly described as magmatically robust.

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Faulting style varies along the length of second-order segments, with faults having larger throws (exceeding 1000 m in places) and wider spacing (>5 km) at segment ends than centres (200–400 m in relief, <5 km spacing) (e.g. Mutter & Karson 1992; Shaw 1992; Shaw & Lin 1993). The change in fault style has been attributed to variations in lithospheric strength along the axis. Thus, large-throw faults develop at the ends of segments where low magma budget and thick axial lithosphere are inferred. And, as at TFs, long-lived detachment faults commonly form at the inside corners of NTOs exhuming lower crustal and upper mantle rocks to the seafloor (Tucholke *et al.* 1998). Studies have shown, however, that long-lived detachment faults can also form in the middle of second-order segments (e.g. Smith *et al.* 2006), and it has been suggested that adjacent detachment faults may link along the axis over a significant portion of a segment (e.g. Reston & Ranero 2011). In some segments, detachment faults can dominate crustal accretion for many millions of years, producing large areas of seafloor covered by extinct detachment faults and exhumed deep-seated rocks (Dick *et al.* 2003; Schroeder *et al.* 2007; MacLeod *et al.* 2008; Smith *et al.* 2008). Escartín *et al.* (2008) estimated that active detachment faults occur along nearly 50% of the MAR axis between 12.5°N and 35°N, which would mean that as much as 25% of new seafloor in this region may be formed by detachment faulting, consistent with earlier estimates (Cannat *et al.* 1995).

Along with increasing water depths, the axial MBA increases by 5–35 mGal as a NTO is approached (Fig. 8; Sempéré *et al.* 1993; Detrick *et al.* 1995; Thibaud *et al.* 1998) supporting the idea of thin crust and thick axial lithosphere at segment ends. An early seismic refraction study at 33°S on the MAR (Tolstoy *et al.* 1993) examined the relationship between the seismically defined crustal thickness and variations in the MBA within a second-order segment, and found that the crust thinned by 2–3 km toward the NTO at the south end of the segment, consistent with the inferences from gravity data. A seismic experiment at 29°N by Wolfe *et al.* (1995) also found thicker crust at the segment centre than at the segment end.

Based on results from a seismic refraction experiment at three adjacent segments south of the Oceanographer TF at 35°N (OH-1, OH-2 and OH-3; Fig. 9a), Hoofft *et al.* (2000) concluded that there were significant variations in crustal thickness (c. 2–3 km) along each segment, with the thickest crust at the centre of each segment and the thinnest crust at the segment ends, consistent with focused magma supply to the segment centre. Interestingly, they found that the average crustal thickness of the three segments was similar (c. 5 km), suggesting

similar total amount of melt supply despite significant differences in axial morphology (e.g. depths along axis, width of the valley floor, etc.) and gravity anomalies. A microseismicity study of OH-1 found that the maximum depth of seismicity (3–4 km) beneath the centre of the segment is anomalously shallow compared to other MAR segments (Barclay *et al.* 2001). The authors speculated that the shallow microearthquakes are related to the thick crust and high crustal temperatures inferred for the centre of the OH-1 segment. Dunn *et al.* (2005) identify a low-velocity anomaly beneath the centre of segment OH-1, which implies high temperatures and perhaps a small amount of melt, and might explain the thicker crust at the segment centre (c. 8.5 km) compared to the segment ends (<4 –5 km).

As a whole this body of work suggests that more melt is supplied to the centre of a slow-spreading second-order segment and that melt supply is limited at the ends of the segments. Whether the melt is supplied vertically from the mantle along the entire axis, primarily to the centre of the axis and distributed along axis at a shallow level, or by a combination of the two remains an area of active research.

Most seismic studies at the MAR have not imaged crustal magma bodies beneath the ridge axis (e.g. Purdy & Detrick 1986; Detrick *et al.* 1990; Canales *et al.* 2000) suggesting that the magma bodies are deep, small and/or ephemeral features (e.g. Sinton & Detrick 1992). Calvert (1995) suggested the presence of a small magma body beneath the segment immediately south of the Kane TF, but the seismic images available were of poor quality. A combined seismic and electromagnetic study along the Reykjanes Ridge (57°45'N) did identify an AML located about 2.5 km beneath a large AVR (Sinha *et al.* 1998). In addition, a large swarm of hydroacoustically and teleseismically recorded earthquakes in 2001 in the Lucky Strike segment near 37°20'N was interpreted by Dziak *et al.* (2002) as a magma/dyke emplacement event into the crust beneath the Lucky Strike volcano, a c. 1 km high edifice in the segment centre; however, there was no along-axis migration of seismicity to indicate lateral propagation of a dyke and subsequent surveys showed no indication of eruption on the seafloor. Subsequently, a seismic reflection study of the Lucky Strike segment revealed a well-defined intra-crustal reflector interpreted as the roof of an AML c. 3 km below the large Lucky Strike volcano (Singh *et al.* 2006a, b).

The spatial and temporal patterns of earthquakes recorded by autonomous hydrophones on the northern MAR (Smith *et al.* 2003; Simão *et al.* 2010) show no simple relationship to second-order segment variables (e.g. length or trend of the segment,

maximum offset of discontinuities or along-axis change in MBA and water depths). Local micro-earthquake experiments have also shown differing trends along axis. For example, Kong *et al.* (1992) observed a progressive shoaling of microearthquakes toward the segment centre at 26°N, which they interpreted as evidence of a recent injection of magma in the region of shallower earthquakes. In contrast, at 29°N, Wolfe *et al.* (1995) did not observe a shallowing of microearthquakes toward the segment centre and suggested that there had not been a recent magma injection there. In general, focal depths of earthquakes in these and other studies (Toomey *et al.* 1988; de Martin *et al.* 2007) range between 3 and 8 km, defining the thickness of the lithosphere at the MAR. It is important to note, however, that the thickness of the seismogenic layer defined by focal depths may reflect short-lived phenomena (e.g. magma movements, fault-slip events) that may not be indicative of the longer-term thickness of the brittle lithosphere.

Geochemical properties of discontinuities and ridge segments

Early reconnaissance sampling of the MAR revealed prominent gradients in isotopic and trace element composition as a function of distance from chemically enriched, near-ridge mantle hotspots, such as Iceland or the Azores (e.g. White & Schilling 1978; Fig. 8b). These regular chemical gradients in source composition often show abrupt changes at TFs such as the Gibbs or Jan Mayen TFs south and north of Iceland (Schilling *et al.* 1983; Dosso *et al.* 1999; Blichert-Toft *et al.* 2005).

The identification of smaller scales of geochemical segmentation along slow-spreading ridges has been problematic for a number of reasons. First, compared to fast-spreading ridges, slow-spreading ridges often show more significant chemical variability over short spatial scales (e.g. Langmuir *et al.* 1992; Batiza 1996; Rubin & Sinton 2007). This local chemical variability was demonstrated, for example, in early studies of the FAMOUS area at c. 36°50'N (Fig. 8; e.g. Langmuir *et al.* 1977) and has been attributed to the prevalence of discrete pockets of enriched mantle (e.g. Michael *et al.* 1994) and to less efficient pooling of chemically distinct melts prior to eruption (e.g. Coogan *et al.* 2000; Rubin & Sinton 2007). Additional challenges in exploring magmatic segmentation at slow spreading rates is that few significant lengths of slow-spreading ridge axes have been sampled at sufficient density and with good age control to fully characterize both inter-segment variabilities.

Despite these complications, sampling in a few areas of the MAR is sufficient to explore relationships between geochemical and tectonic

segmentation and, in particular, the question of central v. multiple magma supply to tectonic segments. In a study of the MAR south of the Kane TF at 24°N, Reynolds & Langmuir (1997) argued for multiple magma supplies based on the observation that melts at the segment ends differed in parental magma composition from those at the segment centre, a finding similar to that reported by Batiza *et al.* (1988) along the southern MAR (see also Niu & Batiza 1994).

Gale *et al.* (2011) studied both intra- and inter-segment variability along the adjacent second-order Menez Gwen and Lucky Strike segments along the MAR. Exploiting the fact that there is a gradient in isotopic composition as a function of distance from the influence of the Azores plume, their results (Fig. 10) document a step function in average isotopic composition across the Pico TF. This observation, combined with the presence of robust central volcanoes, underlain at Lucky Strike by a crustal magma lens (Singh *et al.* 2006a, b) and thicker crust at segment centres, provide general support for the idea that melt supply is focused toward the centre of each segment. Gale *et al.* (2011) further noted, however, that gradients in some trace element ratios within each segment argue in favour of multiple sites of vertical magma supply. Ultimately, these authors concluded that a hybrid model involving both central and lesser amounts of distributed melt supply from depth, followed by shallow along-axis dyking, is likely required to explain the complex geochemical signals.

Because magmas erupted along the MAR often display great chemical diversity resulting from source heterogeneity, even far from hotspots and over spatial scales of a few kilometres or less, it has been difficult to isolate chemical signatures reflecting variations in melting parameters (e.g. Shirey *et al.* 1987; Gale *et al.* 2011, 2013). This makes it problematic to apply the models of primary and secondary magmatic segments developed for fast-spreading ridges (Sinton *et al.* 1991) to the slow-spreading environment.

Ultraslow-spreading ridges

Ultraslow ridges (total spreading rate <20 mm a⁻¹) include the Southwest Indian Ridge (SWIR), the Mohs, Knipovitch and Gakkel Ridges in the North Atlantic and the Mid-Cayman Rise in the Caribbean. These ridges had been largely unexplored until the 1990s, but significant portions are now mapped, both on and off axis (e.g. Mendel *et al.* 1997; Rommevaux-Jestin *et al.* 1997; Grindlay *et al.* 1998; Sauter *et al.* 2001; Okino *et al.* 2002; Cannat *et al.* 2003, 2006; Cochran *et al.* 2003; Dick *et al.*

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Fig. 10. (a–c) The $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ v. latitude for Lucky Strike (circles), Menez Gwen (squares) and KP2–3 segments (diamonds). A sharp jump in isotopic signature occurs between the northernmost end of Lucky Strike and the southernmost end of Menez Gwen and no clear within-segment gradients, in contrast to a more gradual increase seen within segments in the moderately incompatible element ratios (not shown). ‘Central spikes’ in isotopic ratios are also seen within the Lucky Strike segment. From Gale *et al.* (2011).

2003; Michael *et al.* 2003). Because the Mid-Cayman, Mohs and Knipovich ridges are relatively short (<400 km), the range of ultraslow segmentation scales is best investigated on the SWIR and Gakkel ridges. Figures 11 and 12 illustrate characteristics for the eastern part of the SWIR. Until recently, little was known about ultraslow ridges so we begin our discussion with an overview of their geophysical and geological characteristics.

Long ultraslow ridges, such as the Gakkel Ridge (GR) and the SWIR, show a broad range of axial depths and MBA values. As is the case at faster spreading ridges, averages of axial MBA values for long (>200 km) stretches of these ridges are well correlated with average axial depth (Georgen *et al.* 2001; Cochran *et al.* 2003; Michael *et al.* 2003). MBA values typically increase as the ridge axis deepens, indicating higher sub-seafloor

densities either due to changes in average regional crustal thickness or in the average density of the sub-axial mantle, or both. In contrast to what is observed at faster spreading ridges (e.g. at the GSC (Fig. 7) or at the MAR south of the Azores (Fig. 9)), regional axial depth and MBA variations on the SWIR and GR do not as a rule describe large-scale gradients with a deepening of the axis away from hotspots. Instead, average axial depth and MBA at the GR and SWIR vary stepwise over short distances, coincident in some cases with large offset TFs. The Gallieni TF along the SWIR, for example, coincides with a 1000 m deepening of average regional axial depth, and with a *c.* 40 mGal contrast in average regional MBA (Fig. 11). In other cases, however, no surficial tectonic expression is observed other than a NTO (e.g. at 3°E on the GR; Michael *et al.* 2003).

Fig. 11. Variations in ridge properties along the eastern South West Indian Ridge (SWIR). **(a)** Seafloor bathymetry for the SWIR (44–68°E) from the global satellite-derived seafloor topography (Sandwell and Smith, http://topex.ucsd.edu/marine_topo/). Black line marks the ridge axis and its offsets. Location of detailed bathymetry map in Figure 12 is shown by a white rectangle. Transform faults are labelled. In the panels below, shaded vertical bars show the locations of transform faults. **(b)** Basalt geochemistry from Meyzen *et al.* (2005). **(b–e)** Geochemical data, renormalized and culled as described in the caption to Figure 2; data from PetDB (pale crosses) and from Cannat *et al.* (2008; black crosses). **(f)** Mantle Bouguer Anomaly (MBA) modified from Sauter & Cannat (2010). Bars outline areas for which seismic data are available. **(g)** Along-axis depth profile obtained from multibeam bathymetry data (Sauter & Cannat 2010). Segments referred to in the text are numbered.

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Fig. 12. Detailed bathymetry from the SWIR illustrating orthogonal and oblique ridge domains. **(a)** Bathymetry for segments #20 and #21 between the Gauss and Atlantis FZs (see location in Fig. 11). Dashed lines run along axis, in white for oblique domains and in black for orthogonal domains. Grey shaded areas extending off-axis along flow lines from orthogonal domains have thicker than average gravity-derived crust (Mendel *et al.* 2003). **(b)** Along-axis depth profile (detail of Fig. 11g). Note that orthogonal segments are shallower and that the succession of bathymetric lows limits "axial depth segments" that include orthogonal and oblique domains. **(c)** Along-axis MBA profile (detail of Fig. 11f). Note that a large portion of the oblique domain centred at 55.75°E has a flat along-axis MBA profile and that the succession of MBA highs therefore limits "magmatic segments" that mostly correspond to the orthogonal domains.

Active seismic tomographic experiments have been carried out at four locations along the SWIR: near 66°E and 57°E (Minshull & White 1996; Muller *et al.* 1999; Minshull *et al.* 2006); near 51°E (ridge segments 27 and 28; Fig. 11; Zhao *et al.* 2013;

Jian *et al.* 2014); and at 37°E near Marion Island (Seama *et al.* unpublished report, <http://www.inter-ridge.org/de/node/5452>). The seismic structure of the GR has been investigated at 18 locations between 5°W and 60°E, using arrays of geophones

on drifting pack ice as a recording system (Jokat *et al.* 2003; Jokat & Schmidt-Aursch 2007). None of these active seismic surveys have identified negative velocity anomalies that could be interpreted as melt in the crust. The crust is thicker beneath volcanic centres (up to 7 km at SWIR segment 8; Muller *et al.* 1999; Minshull *et al.* 2006). The thickest crust (10.4 km; Jian *et al.* 2014) has been found beneath the centre of the SWIR segment 27 (one of the shallowest ultraslow segments, with a strongly negative MBA; Fig. 11), and the thinnest crust (1.5 km) has been found for the deepest portions (>4500 m) of the GR (Jokat *et al.* 2003; Jokat & Schmidt-Aursch 2007) and of the SWIR east of the Melville Fracture Zone (Muller *et al.* 1999; Fig. 11), which also have more positive MBAs. Widespread ultramafic outcrops in these thin-crust regions indicate that the seismically defined crust in these areas is mostly serpentinized and fractured mantle rocks. Seismically determined crustal thickness, regional axial depth and MBA are thus generally well correlated at ultraslow ridges, as is also found at the slow-spreading MAR.

Ridges spreading at ultraslow rates also share many of the morphologic characteristics of slow-spreading ridges. The ridge axis has an axial valley, as do most ridges with spreading rates $<40 \text{ mm a}^{-1}$ (Small 1998). Most regions of the axis display a succession of along-axis highs and lows with typical half-wavelengths of 30–100 km (Fig. 11; Rommevaux-Jestin *et al.* 1997; Grindlay *et al.* 1998) similar to the second-order segmentation found at slow-spreading ridges (Fig. 8; Sempéré *et al.* 1990). Again, similar to slow ridges, outcrops of serpentinized peridotites are common (Dick 1989; Dick *et al.* 2003; Michael *et al.* 2003; Seyler *et al.* 2003; Hayman *et al.* 2011; Sauter *et al.* 2013). These ultramafic complexes are exhumed by detachment faults that locally bear spreading-parallel 100 m-scaled corrugations (Searle *et al.* 2003; Cannat *et al.* 2009; Hayman *et al.* 2011; Zhao *et al.* 2013) and are similar to corrugated detachment faults at slow ridges (Fig. 9b).

There are, however, characteristics of ultraslow ridge axis morphology that are not found at faster spreading ridges. Most notably, some regions of ultraslow ridges, extending several 100 km along axis ('supersegments' in the terminology proposed by Dick *et al.* 2003) comprise domains of very sparse volcanism that extend up to 80 km along axis. Ultramafic outcrops are ubiquitous in these domains (Dick *et al.* 2003; Michael *et al.* 2003; Sauter *et al.* 2013) and associated with specific mesoscale structures; smooth, flat and uncorrugated seafloor, with sparse hummocky volcanism (Cannat *et al.* 2006; Sauter *et al.* 2013). Regions of the axis that comprise such 'amagmatic' (Dick *et al.* 2003; Michael *et al.* 2003) or 'non volcanic' (Cannat

et al. 2006) ultramafic domains are separated by local regions of strongly localized volcanism, building isolated, widely spaced, and very large axial volcanoes standing >2000 m above the surrounding seafloor (Mendel *et al.* 1997; Cannat *et al.* 1999; Okino *et al.* 2002; Michael *et al.* 2003). Volcanism focused to such a marked extent has not been described at faster spreading ridges. Here we interpret these amagmatic domains with highly focused volcanism as a type of second-order segmentation at ultraslow ridges that will be described in detail below.

Geophysical properties of discontinuities and ridge segments

Transform faults and first-order segments. The spacing of large offset TFs along ultraslow-spreading ridges is typically a few hundred kilometres (Fig. 11), similar to faster spreading ridges. The characteristics of ridge–transform intersections are similar to those of slow-spreading ridges, and there is no regional doming of axial topography or MBA between large-offset transforms. As mentioned before, however, some large transforms coincide with steps in these characteristics. Other large TFs, however, do not mark changes in ridge morphology or geophysical signature (e.g. Indomed TF; Fig. 11).

Second-order discontinuities and segments. There are two primary modes of non-transform bounded segmentation recognized along ultraslow ridges; wide amagmatic domains with highly focused volcanism and orthogonal volcanic domains with oblique basins (Table 1). The large-offset Melville TF (Fig. 11) marks the transition between regions of the SWIR characterized by these different modes. To the west, the succession of axial topographic highs/MBA lows and topographic lows/MBA highs is comparable to the second-order segmentation observed at slow-spreading ridges (Fig. 9). To the east, and extending to the Rodrigues Triple Junction, wide amagmatic domains with highly focused volcanism are found.

Wide amagmatic domains and focused volcanism. In the c. 800 km-long region east of the Melville TF there are three prominent axial volcanoes (numbered 8, 11 and 14 in Fig. 11), each c. 2500 m high and extending c. 60 km along axis, forming spreading-perpendicular volcanic ridges (Mendel *et al.* 1997). The mean along-axis slopes on these volcanoes (up to 90 m km^{-1} ; Cannat *et al.* 1999) are about double those measured west of the Melville TF and along axis in second-order segments of the MAR (Blackman & Forsyth 1991). This contrast in slope suggests distinct mechanisms for creating this topography, and it has

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been proposed that the oceanic crust formed at these volcanoes includes a greater proportion of effusive lava than typical slow-spread crust. The volcanoes rest on seafloor that is very deep (>4500 m) and nearly amagmatic, based on the lack of typical volcanic features such as ridges and cones (Cannat *et al.* 2006) and on dredging of mostly ultramafic rocks (Sauter *et al.* 2013).

The association here of non-volcanic domains with large and widely-spaced axial volcanoes is found in other regions of ultraslow-spreading ridges as well. By their length (several hundred kilometres along axis) these ridge regions would be classified as first-order segments elsewhere, yet they are not as a rule limited by first-order axial discontinuities (i.e. large offset TFs). At the GR non-volcanic domains and focused volcanoes are found between 3°E and 85°E and the limit at 3°E corresponds with a 10 km NTO (Michael *et al.* 2003). Most volcanic centres within these long non-volcanic domains are found between 30°E and 85°E, in the region of the GR that has the slowest spreading rate. In the western part of the SWIR the association of amagmatic domains and large isolated volcanoes is found in the 10–16°E oblique supersegment that is limited to the west by the Shaka TF and to the east by a change in ridge obliquity with no significant ridge offset (Dick *et al.* 2003).

The tall volcanoes are seismically active: swarms of teleseismic events ($M \geq 4.5$) were recorded in 2008 at the SWIR volcanic centre 8 in Figure 11 (Global Seismology Network catalogue) and in 1999 at an axial volcano located at 85°E on the GR (Müller & Jokat 2000; Tolstoy *et al.* 2001). A brief microseismicity study at SWIR volcano 11 (Fig. 11; Yamada *et al.* 2002) suggested hypocentres depths up to 8 km below the seafloor, while a more comprehensive microseismicity study on the Knipovich ridge found maximum hypocentral depths *c.* 20 km beneath the axial basins, and *c.* 8 km beneath the volcanic centres (Jokat *et al.* 2012; Schlindwein *et al.* 2013). The results for axial basins suggest a brittle lithospheric thickness in excess of the 8–10 km that is typically inferred from hypocentre depths at the MAR (Toomey *et al.* 1988; Wolfe *et al.* 1995; de Martin *et al.* 2007). Such a thick axial lithosphere, if confirmed for other ultraslow regions, would mean very cold thermal conditions in the nearly amagmatic domains, and strong thermal contrasts between the volcanic centres and the deeper axial basins.

Dick *et al.* (2003) proposed that the association of non-volcanic domains with large and widely spaced axial volcanoes is typical of extremely low rates ($<12 \text{ mm a}^{-1}$) of effective spreading, either because of ultraslow plate divergence (eastern GR) or because the ridge is highly oblique to the

spreading direction (western SWIR and Knipovich). However, spreading rate does not explain the abrupt change of axial morphologies that occurs with no change in spreading rate across the Melville TF at the SWIR (Fig. 11). There, and also along the GR (Cochran 2008; Goldstein *et al.* 2008), this abrupt transition coincides with significant changes in mantle geochemistry, as discussed below.

Spreading in the amagmatic domains is accommodated by long, axis-parallel detachment faults (Sauter *et al.* 2013). In this tectonic mode of spreading, the axial topographic and MBA profiles are nearly flat (Fig. 11 between volcanoes 8, 11 and 14) and there is no evidence for localized offsets of the ridge axis. In this context, the pronounced along-axis topographic and MBA gradients associated with the volcanoes may define the boundaries of more magmatic second-order spreading segments. Alternatively, several authors have proposed that melt produced in the mantle beneath the adjacent nearly amagmatic domains could be focused for over tens of kilometres toward the volcanoes (Cannat *et al.* 1999, 2003; Okino *et al.* 2002; Sauter *et al.* 2004; Standish *et al.* 2008; Cochran 2008). In this case, the second-order segments in these ultraslow ridge regions would also include part of the amagmatic domains adjacent to axial volcanoes.

Orthogonal volcanic domains and oblique basins. Ultraslow-spreading ridges also comprise regions with more distributed volcanism where domains of smooth amagmatic seafloor are absent or of limited extent, and the axial profile displays successions of topographic highs and lows with typical half-wavelengths of 30 km to 100 km, comparable to the second-order segments of the MAR (Fig. 9). The ends of these segments on ultraslow-spreading ridges typically display elongated basins that trend sub-parallel to the regional orientation of the ridge (Mendel *et al.* 1997; Grindlay *et al.* 1998; Okino *et al.* 2002; Michael *et al.* 2003). This contrasts with segment ends at the slow-spreading MAR, which typically exhibit semi-circular nodal basins adjacent to the discontinuities (Sempéré *et al.* 1990). In addition, on ultraslow-spreading ridges, the elongate basins at segment ends are often oblique to the spreading direction, while the shallower and more volcanically active segment mid-points are often orthogonal to spreading (Fig. 12). In other regions, however, for example locally at the SWIR (Mendel *et al.* 1997; Cannat *et al.* 1999), spreading-perpendicular AVR's are found within the confines of a deep and commonly oblique axial valley. In this configuration, the deeper points of the along-axis profile correspond to the axial valley floor between the volcanic ridges.

The deeper axial regions found between these volcanic zones have a significant width in the

Fig. 13. Sketch showing along-axis section of idealized magmatic spreading segment at fast and slow ridges based on studies summarized in this article. (a) At fast-spreading ridges, rising mantle melts accumulate at the base of the crust (orange–grey) beneath each principal magmatic segment. The crustal magmatic system is composed of a more or less steady state magma lens or sill (red) that is partitioned into finer-scale segments coincident with the finest-scale seafloor segmentation. This shallow magma lens resides above a lower crustal zone of crystal mush and possible lower crustal sills (red lozenges embedded in red–grey). Red arrows in the dyke section indicate trajectories of magma transport during dyking (primarily vertical with minor lateral transport in places). (b) At slow-spreading ridges, strong focusing of mantle melts leads to thick crust/thin axial lithosphere at centre and thin crust/thick axial lithosphere at ends of each principal magmatic segment. Crustal magma bodies are more localized and ephemeral; both vertical and lateral magma transport during dyking may occur. Normal faulting is highly localized in detachments at segment ends and more distributed between several axial valley faults at segment centres. Figure modified from Cannat *et al.* (1995). See text for further discussion.

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Fig. 14. Schematic illustration of ridge segmentation at fast- and slow-spreading ridges, and relationship to mantle melting and ascent. **(a)** At fast-spreading ridges, as well as at intermediate spreading ridges that are magma-rich, the ridge axis is segmented into principal magmatic segments that coincide primarily with third-order tectonic segments (bound on one or both ends by a third-order discontinuity, numbered vertical lines indicate discontinuity order). Each segment receives melt from an upper mantle melt source (thin vertical arrows) that pools at the base of the crust (filled ellipses). Within the crust, melts accumulate within a shallow magma lens (thin line) that is partitioned into smaller segments roughly coincident with the finest-scale tectonic segmentation (fourth-order discontinuities: vertical dash lines). Relatively uniform thickness crust is formed at most segments with modest thinning at some transform faults (vertical arrows labelled '1'). Regional variations in mantle upwelling and melt production at depth (thick arrows) arise due to mantle temperature (shown as undulations in the solidus) and/or compositional variations. **(b)** At slow-spreading ridges, principal magmatic segments coincide with tectonic segments identified as second-order (numbered vertical lines indicate discontinuities of order 1–2). Strong melt focusing (indicated with small subhorizontal to subvertical arrows) at the base of the axial lithosphere (thin dashed line) results in large variations in crustal thickness within segments, with thicker crust typically employed approximately mid-segment and where short-lived crustal magma bodies (filled ellipses in crust) can form.

direction perpendicular to spreading. For example the oblique basin between the shallower and orthogonal ridge domain of segments #20 and #21 (Fig. 12)

has a width of *c.* 50 km in the direction perpendicular to spreading and therefore accommodates plate divergence over a significant length of the plate

boundary. This has led to the idea that these wide and deep domains of ultraslow-ridges represent a new class of spreading segments (Dick *et al.* 2003). Other workers have interpreted these domains as wide axial discontinuities separating short volcanic segments (Mendel *et al.* 1997). While these interpretations may both be valid, one could argue (e.g. Cochran 2008) that a true magmatic spreading segment (as opposed to more specific definitions such as a tectonic or volcanic segment) should have its own melt supply system. Most workers to date have hypothesized that melt produced in the mantle beneath these deeper portions of the axis is then focused to the more volcanically robust orthogonal regions, based on geophysical (Rommevaux-Jestin *et al.* 1997; Grindlay *et al.* 1998; Cannat *et al.* 1999, 2003; Okino *et al.* 2002; Jokat & Schmidt-Aursch 2007) and geochemical (Standish *et al.* 2008) arguments as described below.

Geochemical properties of discontinuities and ridge segments

No ultraslow spreading ridge segments have been sampled at densities sufficient to explore intra- or inter-segment chemical variations, like those described for the slow-spreading northern MAR (Gale *et al.* 2011). The SWIR is currently the most densely sampled ultraslow spreading ridge (Fig. 11), but even here data are only sufficient to describe regional variations and their possible correspondence to first-order segmentation. Based on present sampling of the SWIR from 44°W to 68°W, magmatic segmentation defined by changes in isotopic composition appears to coincide with some first-order discontinuities, such as the Melville and Atlantis TFs (Fig. 11) (Meyzen *et al.* 2005). Some intervening first-order segments appear to erupt magmas of relatively uniform isotopic and trace element composition (e.g. segments spanning the Atlantis, Novara and Melville transforms), while other segments show large-amplitude isotopic variations over small scales, independent of ridge segmentation.

With respect to variations in extents of melting, average $\text{Na}_{8.0}$ values increase from 44°W to 68°W, consistent with the *c.* 3000 m decrease in average axial depth, suggesting decreasing extents of melting (Fig. 11) (Cannat *et al.* 2008). In the more densely sampled area between the Atlantis and Melville TFs, $\text{Na}_{8.0}$ values grade smoothly through the two first-order segments, suggesting that despite the isotopic distinctions between these first-order offsets (Meyzen *et al.* 2005), variations in mantle melting occur independent of tectonic segmentation. Finally, apart from the eruption of more fractionated lavas in the vicinity of some

transforms, variations in extents of fractionation (MgO wt%) show little systematic inter- or intra-segment relationships. Further sampling of ultraslow spreading centres at scales and densities sufficient to explore both intra- and inter-segment variability is required to understand the extent to which ultraslow spreading centres represent an end-member in a continuum as a function of spreading rate, or display entirely unique geochemical characteristics.

Synthesis and discussion

Geophysical and geochemical studies conducted since the 1980s confirm early inferences that the tectonic segmentation of the MOR defined by transform and non-transform discontinuities of varying sizes is closely linked to segmentation in the magmatic system that feeds the ridge. Variations in a range of ridge properties are observed at all spreading rates that indicate tectonic offsets of the ridge axis coincide with disruptions in the underlying magma supply system. Furthermore, ridge segments at fast- and intermediate-spreading ridges, separated in places by discontinuities with very small offsets of <2–3 km, often display distinct properties indicating they are fed by different magmatic systems. The notion that the global MOR is composed of a series of magmatic spreading segments, each with its own magma plumbing system extending into the underlying asthenosphere is generally supported.

In Figure 13, we summarize the main features of what we refer to as a 'principal magmatic segment' defined as the length of ridge fed by a discrete melt accumulation in the shallow asthenosphere at fast- and slow-spreading ridges. While the model of a slow-spreading principal magmatic segment is well supported by a diverse suite of observations of the MAR over the past two decades, the model of a fast-spreading principal magmatic segment is more speculative and is based on recent observations from a limited portion of the northern EPR. Figure 14 summarizes inter-segment variations and implications for mantle upwelling and melting for comparison with Figure 1. In the discussion that follows, we first describe the general characteristics of principal magmatic segments and their relationship to mantle processes. We then summarize insights gained from current observations regarding early ideas of spreading cells, hierarchical segmentation, central magma supply beneath ridge segments, and the relationship between geochemical and geophysical indicators of magmatic segmentation. This discussion builds upon the numerous studies cited in the preceding sections: references are included below to support concepts

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not previously introduced and for the historical discussion of the later section on the concept of spreading cells.

Melt distribution and structure of principal magmatic spreading segments

Beneath fast-spreading ridges (Figs 13a & 14a), current geophysical and geological observations suggest that rising mantle melts accumulate and pool at the base of the crust forming discrete melt reservoirs in the uppermost mantle at intervals roughly coincident with third-order tectonic segments. Some focusing of melts within the upper mantle is required to form these localized regions of melt pooling. Within the crust, magma lenses or sills consisting of a few percent to 100% melt exist as approximately steady state bodies at the top of a lower crustal zone of crystal mush and may also form at mid- and lower-crustal levels within this mush zone. Magma bodies in the crust are further partitioned into lenses that often coincide with the finer (fourth-order) seafloor segmentation of the active eruptive fissure zone. At fast spreading rates, crustal thickness is approximately uniform within these principal magmatic segments. The primary characteristics of principal magmatic segments at fast-spreading ridges depicted in Figure 13a may also be relevant for the magmatically robust portions of intermediate-spreading ridges, although the available data are more limited.

Slow-spreading ridges differ markedly from fast-spreading ridges in crustal and lithospheric structure and in the distribution of melt within each magmatic segment (Figs 13b & 14b). Strong focusing of mantle melts within each principal magmatic spreading segment (defined by the second-order segmentation) leads to large variations in crustal thickness (2–3 km), with thick crust and thin axial lithosphere at segment centres, and thin crust and thick axial lithosphere at segment ends. Crustal magma bodies may form within a segment but are more localized and ephemeral than at fast-spreading ridges: melt mush and melt lenses may develop at various levels in the crust and upper mantle lithosphere along the segment but are concentrated primarily near segment centres. At magma-rich segment centres, the magmatic crust is more continuous and layered, but toward the magma-poor segment ends, the magmatic crust becomes progressively thinner and more discontinuous, and lower crustal and upper mantle rocks may be exhumed by large-offset detachment faults. Detachment faults can reach horizontal offsets of many tens of kilometres compared to the shorter-lived normal faults commonly formed near segment centres that only extend for a few hundred metres, before being abandoned as a new fault

initiates closer to the axis. The more magmatic portions of ultraslow-spreading ridges share the primary features of principal magmatic segments as slow-spreading ridges. The morphological and geological characteristics of the discontinuities that separate magmatic segments at these ridges are diverse and often tectonically complex, particularly at ultraslow ridges.

The concept of spreading cells

The original concept of a spreading cell developed in the early 1980s from studies of magnetic anomalies of old Atlantic crust was that the MOR is subdivided into quasi-regular segments, each with its own mantle melt supply system at depth (Schouten & White 1980; Schouten & Klitgord 1982). Although the offsets between segments (believed at the time to be TFs) could change length and even disappear, the segments persisted as long-lived coherent units. As detailed bathymetric data from the MAR were acquired, it became apparent that this spreading cell segmentation corresponded with what was later classified as the second-order segmentation of the ridge bounded by NTOs (Schouten *et al.* 1985; Sempéré *et al.* 1990). The OSCs discovered at fast-spreading ridges were believed to be analogous features bounding similar long-lived spreading cells at the EPR (Macdonald *et al.* 1984; Schouten *et al.* 1985). Interpreted spreading cells along the MAR were *c.* 50 km in length (Schouten *et al.* 1985) and somewhat longer along the EPR (65 km from Schouten *et al.* (1985); 145 km from Macdonald *et al.* (1992)), although as more of the EPR was mapped in detail, it became apparent that the interpreted spreading cell segmentation did not exhibit the same regularity as along the MAR. Furthermore, a growing body of evidence at fast-spreading ridges argued against a centralized magmatic plumbing system beneath the second-order segments bounded by OSCs.

Observations from many ridge areas surveyed over the 30 years since these early studies reveal a quasi-regular segmentation of the ridge into principal magmatic segments that is analogous to the original notion of spreading cells. At slow-spreading ridges, principal magmatic segments (Fig. 13b) coincide with the second-order segmentation and correspond with the spreading cells as originally interpreted by Schouten and colleagues. Regional-scale studies of the MAR using the Sandwell & Smith (1997) global bathymetry compilation reveal a range of 20–100 km and an average length of 50 km for this segmentation (Briais & Rabinowicz 2002), consistent with original estimates based on studies of more limited portions of the MAR. Ultraslow-spreading ridges include long melt-poor

sections with volcanic centres that are separated by larger distances (200 km spacing or more) with non-volcanic domains between them. Other, probably more magmatic regions of ultraslow ridges, however, have second-order segments that are c. 30–100 km long, similar to slow-spreading ridges.

At fast-spreading centres, principal magmatic segments appear to coincide with the third-order tectonic segments of the ridge (Fig. 13a) and not the larger-scale second-order segments originally interpreted as spreading cells. These segments range in length from 20 to 90 km (average 30 km for northern EPR), similar to these segments at slow spreading rates. Data from the hotspot-dominated portions of the intermediate-spreading GSC suggest distinct spreading segments with dimensions of 30–120 km, also much shorter than the second-order segmentation of this ridge. At the intermediate spreading SEIR, second- and some third-order segments may correspond with principal magmatic segments. The regional perspective provided by the Briais & Rabinowicz (2002) study of satellite-derived bathymetry indicates an average length of 50 km for the primary segmentation of intermediate ridges including the SEIR.

Hence a pattern emerges of principal magmatic segments that are similar in length (30–50 km on average) at all spreading rates. The characteristics of these segments inferred from the modern data are distinct from the spreading cells originally proposed by Schouten *et al.* (1985) and others in that each is linked to a distinct magma plumbing system in the shallowest asthenospheric mantle (e.g. at >6 km to >12 km depending on spreading rate) although the observations of punctuated melt centres at fast-spreading ridges likely require deeper diapiric upwelling of melts. At fast-spreading ridges the principal magmatic segments inferred from the modern data are present on a finer scale than the spreading cells originally proposed for the EPR, and these cells may be shorter lived at these spreading rates.

Chemical observations and inferences regarding magmatic segmentation

Geochemical observations from fast-spreading ridges indicate that mantle source compositions and extents of melting typically vary over length scales that are longer than the tectonic segmentation of the ridge defined by TFs and NTOs. However, where abrupt changes in lava chemistry occur that indicate significant differences in source composition or melting processes, they usually coincide with first-, second- or third-order tectonic discontinuities. Yet many tectonic discontinuities show no change in geochemical parameters. This suggests that where chemical discontinuities coincide

with offsets along fast-spreading ridges, they may arise from shallow upper mantle melt focusing and crustal melt distribution processes, independent of regional variations in mantle source composition and extents of melting.

Along slow-spreading ridges, magma compositions often vary widely over short spatial scales most likely due to inefficient melt pooling and possible prevalence of small-scale mantle heterogeneities. In the few places where seafloor sampling is sufficient to characterize inter-segment variations, geochemical data show some evidence of differences in composition between adjacent spreading segments, superimposed upon long-wavelength gradients that cross multiple first- and second-order segment boundaries. As for fast-spreading ridges, this is consistent with the notion of upper mantle focusing of melt to individual magmatic spreading segments.

Thus, while there is some correspondence between geochemical and geophysical segmentation of the ridge over a range of spreading rates, existing data suggest these relationships are complex. To be sure, on a global scale, there are strong correlations between geophysical determinations of the volume of melt delivered to the ridge (leading to a given crustal thickness and axial depth) and geochemical determinations of the volume of melt produced by mantle melting (leading to a given melt composition and inferred extent of mantle melting). But in our effort to study axial variations on smaller scales, analysis of geophysical and geochemical data often yields conflicting results. There are likely a number of reasons for this current mismatch. First, much of the geophysical data present a time-averaged view of crustal construction, while geochemical analyses of individual lavas represent a snapshot in time of only a small fraction of the melt that over time will build the crust. Second, geochemical studies show that diverse melt compositions are produced during mantle melting that may pool to greater or lesser extents in different settings. Third, in addition to regional gradients in mantle source composition, the existence of smaller-scale heterogeneities presents significant challenges in our effort to characterize along-axis variations in melting parameters. Yet despite these complexities, there are aspects of magma supply that geochemical studies are uniquely suited to address, because when deciphered, melt compositions reveal their distinctive histories that can be used to track magma transport and distribution.

Is segmentation hierarchical?

The notion of a hierarchy of nested scales of magmatic segments defined by distinct classes of

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ridge axis discontinuities was proposed in the late 1980s and has been an influential concept in MOR research. However, the perspective gained from observations of many MOR regions indicates that the applicability of this hierarchical model, with first-order TF bounded segments reflecting deeper mantle upwelling and melting processes, partitioned into smaller segments that share the deep plumbing (Fig. 1), is limited. The prominent along-axis topographic arches present along slow- to ultraslow-spreading ridges and some intermediate ridges, which are attributed to along-axis variations in magma supply, occur on the 30–50 km length scales of principal magmatic segments and not on the longer length scales of the first-order segments (Figs 5, 8, 10). While some TFs do mark major boundaries in geochemical properties, indicating distinct mantle source compositions (e.g. the Romanche and Oceanographer TF along the MAR), such geochemical boundaries also occur at smaller offsets (e.g. the 20°40'S OSC on the EPR (Fig. 4), or the 92°40'W offset on the GSC (Fig. 7)). Changes in ridge properties, such as axial depth, axial gravity anomalies, crustal thickness and lava chemistry, approaching and near TFs, are often comparable to those observed at smaller offsets (e.g. compare north of the Clipperton TF with north of 20°40'S OSC; Figs 2 & 4). Even very small ridge axis discontinuities can be major regional boundaries in a range of ridge properties (e.g. 92°40'W on the GSC; Fig. 7).

There is evidence, however, for larger-scale 'tectonic corridors' that span multiple TF- and NTO-bounded segments at intermediate and slow ridges and 'super-segments' at ultraslow ridges that may be significant for deeper mantle processes. The obliquity of the ridge relative to spreading direction is also believed to contribute to regional variations at these length scales. In addition, the notion of hierarchical nested scales of segmentation does provide a useful framework for understanding the highest order segmentation at fast- and possibly intermediate-spreading ridges. At these ridges, third-order segments, which appear to coincide with segmentation of melt in the shallowest mantle, are subdivided into smaller fourth-order segments that may primarily reflect partitioning of melt at shallower levels within the crust (Fig. 13a).

Central magma supply beneath spreading segments

At slow and ultraslow ridges, modern observations confirm early inferences that mantle melt is highly focused beneath the centres of what are now commonly referred to as second-order ridge segments, leading to thicker crust and thinner axial lithosphere within these regions (Fig. 13b). While these

observations are consistent with the notion of centralized melt supply, this melt focusing may result from processes in the shallow mantle and does not necessarily require melt diapirs rising from depth as originally proposed by Whitehead, Schouten and others. Large spatial gradients in the thickness of the axial lithosphere are expected both across and along axis at slow and ultraslow ridges (Figs 13b & 14b) that may act to efficiently focus mantle melts that migrate along the base of the lithosphere (e.g. Sparks & Parmentier 1991; Magde & Sparks 1997). Furthermore, strong positive feedback between pooling of mantle melts and lithospheric structure is expected, and magmatic segmentation could be primarily controlled by spatial variations in lithospheric thickness. While the observations do not rule out some deeper (e.g. 30–50 km) organization of mantle upwelling and melting, melt focusing at slow- and ultraslow-spreading ridges could thus be largely controlled by the base of the lithosphere (e.g. Montesi *et al.* 2011).

At fast-spreading ridges, however, lithospheric controls are less likely to account for the local accumulations of mantle melts that have been detected beneath third-order segments by Toomey and colleagues (Fig. 13a). Here, the base of the lithosphere is within or at the base of the crust, along-axis variations in crustal thickness are modest, and hence melt pooling guided by variations in lithospheric thickness should be minimal. At these ridges, variations in asthenospheric mantle upwelling and/or melt segregation processes are likely required to account for the presence of punctuated melt accumulations.

Questions for future work

Current observations suggest that a primary tectonic and magmatic segmentation of the MOR system exists arising from the relatively regular distribution of mantle melt within the shallow asthenosphere with a characteristic length scale of 30–50 km. This inference is derived from multi-disciplinary studies that have been conducted in a few select regions: clearly, future studies at a spectrum of sites will be required to test this interpretation. Critical areas for future study pertaining to this notion of principal magmatic segments at all spreading rates include: 3D studies of the segment-scale distribution of melt in the asthenospheric mantle beneath ridges over a range of spreading rates; detailed multi-segment geochemical and geological studies along slow- and ultraslow-spreading ridges; and new modelling studies to explore the origin of the quasi-regular segmentation at all spreading rates and the significance of deeper mantle processes in their development.

Other important questions not touched upon in this review include: how segments initiate (e.g. Bird & Naar 1994; Cormier *et al.* 1996); the role of lateral magma transport in the crust and mantle (e.g. Macdonald *et al.* 1984; Fox *et al.* 1995; Batiza 1996); the contribution of small upper mantle melt anomalies to segmentation (e.g. Gomez & Briaies 2000); how discontinuities evolve and grow, in particular the relationship between second- and third-order offsets at fast and intermediate ridges (e.g. Carbotte & Macdonald 1992; Cormier *et al.* 1996; Canales *et al.* 2002); and what drives the migration of these discontinuities (e.g. Macdonald *et al.* 1984, 1988; Schouten *et al.* 1987; Tucholke & Schouten 1988; Lonsdale 1989, 1994; Cormier 1997). Some of the critical needs pertaining to these areas of active research include: integrated geophysical and geochemical studies that target the ridge flanks to understand the processes that cause magmatic segments to initiate, migrate and evolve through time; long-term multi-segment-scale microseismicity monitoring studies with OBS arrays to detect and characterize magma transport in the crust; and geochemical and geochronological studies of both the ridge axis and flanks to explore changes in melt composition as ridge segments evolve.

Thanks to many colleagues for stimulating discussions and for providing data, including Ken Macdonald, Mike Perfit, Charlie Langmuir, Pablo Canales, Milene Cormier, Javier Escartin and Daniel Sauter. We also thank Cindy Ebinger and Charlie Langmuir for their helpful reviews which contributed to a much improved paper.

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