Hyperextended continental margins—Knowns and unknowns

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In this issue of Geology, Tugend et al. (2015, p. 15) discuss rift geometries and hyperextension in the Bay of Biscay–Parentis (BBP) area. This is a well-defined propagating rift/ocean (e.g., Sibuet et al., 2012), with crustal architecture revealing a succession of zones typical of magma-poor margins: (1) limited crustal stretching, (2) hyperextension, (3) exhumed mantle, and (4) ultraslow or normal oceanic crust (cf. Péron-Pinvidic and Manatschal, 2009). The zones probably formed as the rift propagated, a common mode of continental breakup (e.g., the South Atlantic, Heine et al., 2013; Central Atlantic, Kneller and Johnson, 2011, and North Atlantic oceans; Sibuet et al., 2012). Tugend et al. show how this margin evolved as partitioned transtensional domains along the nascent plate boundary and, perhaps most significantly, describe how hyperextension along the boundary localized deformation during later plate convergence.

Hyperextension is defined as stretching of the crust such that the lower and upper crust become coupled and embrittled, allowing major faults to penetrate to the mantle, leading to partial hydration (serpentinization) of the uppermost mantle, as observed in the thickened blocks in the Swiss Alps (e.g., Manatschal, 2004). Hyperextension is documented worldwide, e.g., in the South Atlantic (Contrucci et al., 2004), off southern Australia (Díireen et al., 2007) and in the Red Sea (Cochran and Karner, 2007), and requires stretching by a factor of 3–4, with crustal thinning to ~8 km or less (e.g., Pérez-Gussinyé and Reston, 2001). Off Iberia-Newfoundland and in the Labrador Sea, fully exhumed serpentinitized mantle forms 50–100 km belts along the hyperextended basins and (“Penrose”) oceanic crust with linear magnetic anomalies (Fig. 1).

The Iberian and Newfoundland conjugate margins are testing grounds for the concept of hyperextension, with exhumed mantle rocks, serpentinitized peridotites of continental affinity, sampled at Ocean Drilling Program (ODP) sites (e.g., Whitmarsh and Miles, 1995; Tucholke et al., 2004). Fault blocks have been drilled and correlated with seismic data on the Galicia Bank and seaward (e.g., Tucholke et al., 2007; Ranero and Pérez-Gussinyé, 2010). ‘Necking’, the process whereby the lithosphere thins through time and across a basin, is controversial in these margins. Conjugate margins are commonly asymmetric, one margin characterized by gradual thinning with a wide transition between continent and ocean, the other by more abrupt thinning and a narrower transition (e.g., Hopper et al., 2004). Both types, however, show a discrepancy between overall crustal thinning and stretching as derived from seismically observable brittle faults. Models to explain this discrepancy include large-scale crustal detachments (Lister et al., 1991), depth-dependent (coaxial) stretching (i.e., differential extension of upper brittle and lower ductile crust), and masking of extensional events by polyphase faulting (Reston, 2007). Recent models explain broad hyperextended margins such as offshore Galicia or Angola in terms of sequential upper crustal faulting, with the extension focus migrating oceanward, and with upper crustal brittle faulting balanced by lower crustal flow (e.g., Ranero and Pérez-Gussinyé, 2010). Higher resolution seismic grids are needed to decide which model, or hybrid of models, is valid.

Early Cretaceous breakup along the Galicia margin and in the Bay of Biscay may represent the southern end-member of a chain of hyperextended basins along the north Atlantic (e.g., Lundin and Doré, 2011), all characterized by highly thinned crust. All failed to achieve full oceanic status, subsided rapidly and accumulated sedimentary infill (10 km or more in the Møre Basin). Sub-crustal bodies in some basins have been identified as serpentinitized mantle (e.g., O’Reilly et al., 1996, Reyinnesson et al., 2010), but this can be difficult to identify from velocity structure alone: P-wave to S-wave velocity ratios (Vp/Vs) can be used as a lithology indicator, but at a Vp range of 7.0–7.5 km/s there is significant overlap with gabbroic rocks (“underplate”), and other potential sub-crustal bodies (e.g., Mjelde et al., 2009).

The most intriguing aspect of hyperextended margins may be what differentiates them from volcanic passive margins (Fig. 1 and Table 1; see also Franke, 2013). Hyperextended margins result from slow extension rates and are magma-poor, but it is not clear why. The assumption of slow extension rates is based on initial sea-floor spreading half-rates in the order of 10 mm/yr (e.g., Sibuet et al., 2004). Magma-rich margins, in contrast, are more sharply defined and co-associated with initial spreading half-rates of ~25–30 mm/yr (e.g., Schreckenberger et al., 2002; Hopper et al., 2003). Extension rates immediately prior to break-up are difficult to quantify in hyperextended margins, which provide few clues as to the duration of their evolution. Even the well-studied Iberia margin provides limited constraints, since both phases are short and on structural highs rather than in syn-rift wedges, and rift successions tend to be too deeply buried to interest the

### TABLE 1. FACTORS DIFFERENTIATING HYPEREXTENDED (MAGMA-POOR) AND VOLCANIC PASSIVE MARGINS

<table>
<thead>
<tr>
<th>Hyperextended Margin</th>
<th>Volcanic Margin</th>
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</thead>
<tbody>
<tr>
<td>Magma-poor (mantle melt – 10%)</td>
<td>Narrow necking zone</td>
</tr>
<tr>
<td>Wide necking zone</td>
<td>Rapid extension rate</td>
</tr>
<tr>
<td>Slow extension rate</td>
<td>Seaward-dipping reflectors (SDRs)</td>
</tr>
<tr>
<td>Serpentinitized mantle</td>
<td>Clear Moho</td>
</tr>
<tr>
<td>No Moho in outer (exhumed) zone</td>
<td>Subaerial breakup</td>
</tr>
<tr>
<td>Deep-water breakup</td>
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Figure 1. Schematic structure of hyperextended (magma-poor) and magma-rich passive margins.
petroleum industry. Exhumed mantle is particularly challenging, because it cannot be isotopically dated and does not show linear magnetic anomalies. Differences in magmatism between the two margin types are also a rich topic for debate. Hypertended margins are not amagmatic: exhumed mantle contains up to ~12% infiltrated melt in the Swiss Alps and Iberian margin (Müntener et al., 2010). The degree of melting varies from ~10% to 100% between magma-poor and magma-rich margins. The latter (e.g., the northeast Atlantic) are attributed by most workers to elevated mantle temperatures associated with a plume (e.g., Smallwood and White, 2002). It is commonly assumed that the rapid continental break-up and the initial spreading are caused by this additional heat source, absent in hyper-extended margins. Alternatively, reversing cause and effect, the amount of magmatism may be a function of the spreading rate, correlated with distance from the plate tectonic pole of rotation, so that slower separation rates and magma-poor margins occur closer to the pole (Lundin et al., 2014).

Hypertended magma-poor margins will probably react differently to ocean closure than stronger, thicker magma-rich margins. Hypertended lithosphere is deformation-prone due to extreme crustal thinning and partial replacement of peridotite by rheologically weak serpentinite, and exhumed mantle will be even weaker. Such zones may have focused compressional deformation on the North Atlantic margin (Lundin and Doré, 2011) and such weak elements may become important in localizing subduction when oceans close (Tugend et al.). Thus, hypertended margins may have significant roles to play at critical stages of the Wilson Cycle, the process whereby oceans open and close along broadly similar lines during supercontinent break up (Wilson, 1966).

REFERENCES CITED


