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Wilson cycle passive margins: Control of orogenic inheritance on continental breakup

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ABSTRACT

Rifts and passive margins often develop along old suture zones where colliding continents merged during earlier phases of the Wilson cycle. For example, the North Atlantic formed after continental break-up along sutures formed during the Caledonian and Variscan orogenies. Even though such tectonic inheritance is generally appreciated, causative physical mechanisms that affect the localization and evolution of rifts and passive margins are not well understood.

We use thermo-mechanical modeling to assess the role of orogenic structures during rifting and continental breakup. Such inherited structures include: 1) Thickened crust, 2) eclogitized oceanic crust emplaced in the mantle lithosphere, and 3) mantle wedge of hydrated peridotite (serpentinite).

Our models indicate that the presence of inherited structures not only defines the location of rifting upon extension, but also imposes a control on their structural and magmatic evolution. For example, rifts developing in thin initial crust can preserve large amounts of orogenic serpentinite. This facilitates rapid continental breakup, exhumation of hydrated mantle prior to the onset of magmatism. On the contrary, rifts in thicker crust develop more focused thinning in the mantle lithosphere rather than in the crust, and continental breakup is therefore preceded by magmatism. This implies that whether passive margins become magma-poor or magma-rich, respectively, is a function of pre-rift orogenic properties.

The models show that structures of orogenic eclogite and hydrated mantle are partially preserved during rifting and are emplaced either at the base of the thinned crust or within the lithospheric mantle as dipping structures. The former provides an alternative interpretation of numerous observations of 'lower crustal bodies' which are often regarded as igneous bodies. The latter is consistent with dipping sub-Moho reflectors often observed in passive margins.

1. Introduction

Passive margins form the transition between oceanic and continental lithosphere and are the result of rifting and continental break-up. These processes represent one phase of the Wilson cycle (Wilson, 1966; Dewey and Spall, 1975) where continents recurrently disintegrate and reassemble and oceanic crust and lithosphere form, subduct and recycle, a process that is likely to have been ongoing for around 3 Ga (Cawood, 2006; Shirey and Richardson, 2011). The present-day margins of the Central and North Atlantic are a manifestation of at least two complete Wilson cycles: The supercontinent of Rodinia was assembled at the end of the Mesoproterozoic Grenvillian orogeny (Piper, 2000; Thomas, 2006), which after break-up was divided by the newly forming Iapetus Ocean. The closure of the Iapetus Ocean leads to the Paleozoic Caledonian–Acadian orogeny in the North Atlantic region. This continent–continent collision involved the continents Laurentia, Baltica, Avalonia and further smaller continental fragments and terranes (Van Staal et al., 1998; Leslie et al., 2008) and formed a coherent, Himalaya-type mountain range of at least 3000 km length and 1000 km width (Roberts, 2003; Gee et al., 2008). The subsequent Late Paleozoic Variscan orogeny assembled large parts of present day Central and Southern Europe partly overprinting the Caledonian Orogen in the Appalachians (Stamfli and Kozur, 2006). Most recently, Mesozoic and Cenozoic rifting caused the breakup of the supercontinent Pangea and the formation of the North Atlantic and passive margins (Skogseid et al., 2000).

Remnant structures and lineaments of earlier mountain-building events are generally strike-parallel and mimic the present-day North Atlantic margins, implying ancestral control of the older orogens (Williams, 1995 and references therein). In a recent review by Buiter and Torsvik (2014) it is concluded that continental breakup generally occurs along former collision zones irrespective of their age. In another recent paper, Schiffer et al. (2015b) demonstrated that the plate tectonic

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evolution of the North Atlantic region is likely controlled by inherited orogenic structures formed during the Caledonian.

Despite general agreement that orogenic structures can control the location of extension (e.g. Dunbar and Sawyer, 1989; Williams, 1995), few studies quantitatively address possible implications of orogenic inheritance during the formation and evolution of passive margins. In this paper we explore such implications using numerical modeling of the rift-to-drift evolution that forms as a result of extension of lithosphere with rheological inhomogeneities inherited from the time of suturing and orogeny. We investigate what influence such structures have on structural properties and the evolution of passive margins. These include the asymmetry of conjugate margins, the formation of lower crustal bodies (LCBs) of different shapes, hyperextension, formation of blocks of thinned continental crust (Péron-Pinvidic and Manatschal, 2010; Lundin and Doré, 2011) and the rate and timing of melt generation. Before describing our modeling procedure and our results, we first present a general summary of such properties of passive margins.

2. Magma-rich vs. magma-poor margins

Passive margins accommodate the transitional thinning of the continental crust towards the oceanic sea floor (Fig. 1) and are often discriminated in terms of the volume of igneous rocks emplaced during thinning of the continental crust. The Iberia–Newfoundland conjugate margins (Fig. 1e–g) represent a much cited example of a conjugate pair of magma-poor margins. Here, igneous rocks are absent or sparse (Minshull et al., 1998), and blocks of thinned continental crust are in tectonic contact with exhumed and serpentinized mantle ('extensional allochthons'; see Wernicke, 1981; Froitzheim and Manatschal, 1996; Péron-Pinvidic and Manatschal, 2010). The exhumed mantle forms a distal zone that gradually transitions to 'normal' oceanic crust (Boillot et al., 1980; Bown and White, 1994; Reston and Morgan, 2004; Péron-Pinvidic and Manatschal, 2010). In contrast, magma-rich passive margins, such as the Greenland–Scotland/Norway margins (Fig. 1a–c), are associated with abundant igneous rocks in the form of subaerial and submarine extrusive basalt flows (Mutter et al., 1982; Tegner et al., 1998) or lower crustal intrusions (White, 1992; Coffin and Eldholm, 1994; Menzies, 2002). The total thickness of igneous rocks is, at places, reported to be several times the thickness of normal oceanic crust (Kelemen and Holbrook, 1995; Korenaga et al., 2000).

An apparent difference between magma-rich and magma-poor margins is that the onset of melt production tends to precede complete breakup of the continental lithosphere during the formation of magma-rich margins, forming igneous intrusions in and flood basalts on top of continental crust. The opposite seems to be the case for magma-poor margins (Manatschal et al., 2015), where no or few igneous products are emplaced in or on the thinned continental crust. Igneous rocks instead appear to have formed after the crust was completely thinned and mature oceanic spreading onset.

A number of models have been invoked to account for the differences between the two types of margins. For example, Reston and Morgan (2004) suggested that magma-poor margins can be explained by a ~100 °C reduction of mantle temperature relative to the 'normal' ~1300 °C, typically required to produce 6–7 km oceanic crust (e.g. Bown and White, 1994). Similarly, to explain magma-rich margins, hotter-than-normal mantle temperatures have been invoked, such as plumes (White and McKenzie, 1989). The plume-origin of Large Igneous Provinces and associated magma-rich margins has been disputed (Holbrook et al., 2001; Foulger et al., 2005; Foulger and Jurdy, 2007). Alternative mechanisms include small-scale convection (Mutter et al., 1988; King and Anderson, 1998; Boutilier and Keen, 1999), time/depth-dependent thinning of the lithosphere (van Wijk et al., 2001), enhanced melt source fertility (Korenaga and Kelemen, 2000; Korenaga, 2004) and water in the mantle source (Jamtveit et al., 2001).

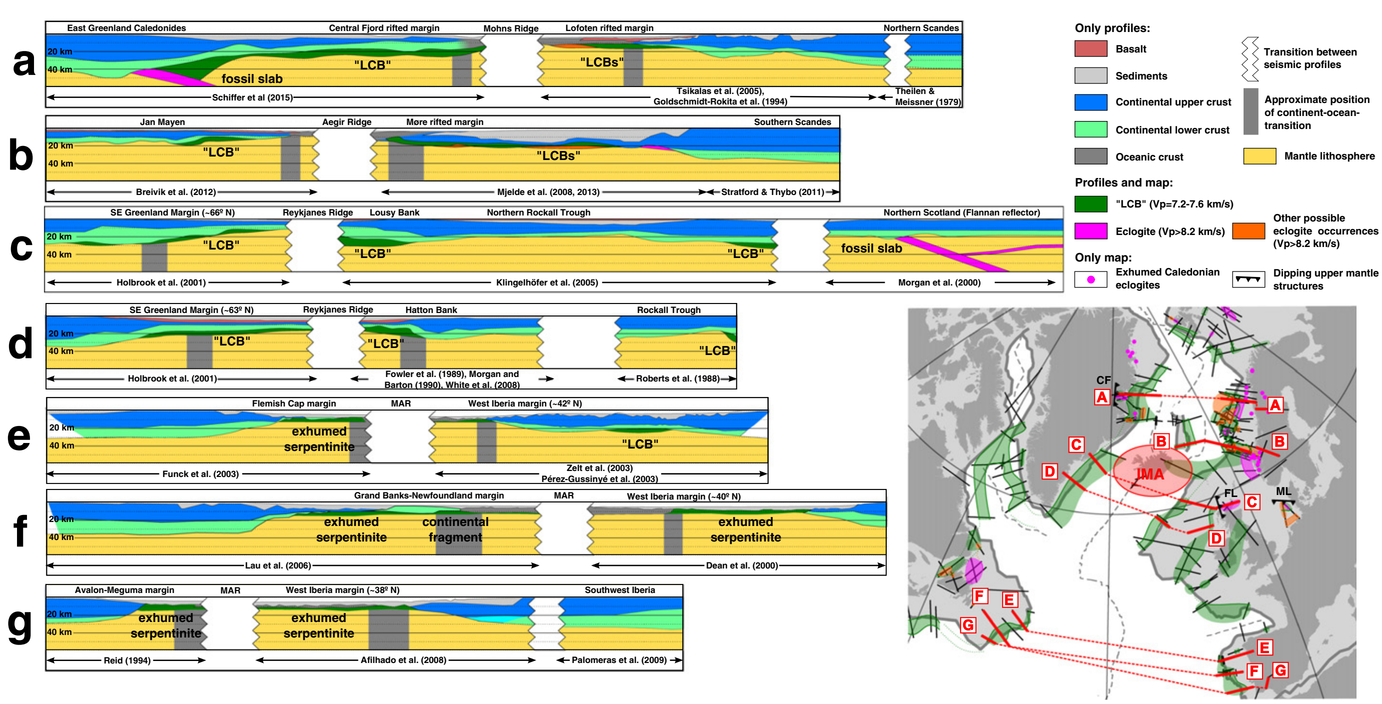

Fig. 1. North Atlantic Crustal Transects from the East Greenland–Norwegian (a–b), East Greenland–British (c–d) and Iberia–Newfoundland margins (e–g). The transects show sediments (gray), basalts (red), upper and lower crust (dark and light blue) as well as lower crustal bodies of different kind. Magenta shows previously interpreted eclogite bodies; orange shows other areas with Vp > 8.2 km/s at the base of the crust, which may be attributed to the presence of eclogite; green shows lower crustal bodies consistent with mafic underplating or serpentinite bodies. The map illustrates the position of the transects (red) and wide-angle seismic lines (black lines) in the North Atlantic. Magenta, orange and green are the same as for the transects. IMA — Iceland Melt Anomaly. CF — Central Fjord structure. ML — Mona Lisa structure. FL — Flannan structure.