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## Development of volcanic passive margins: Three-dimensional laboratory models

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[1] Continental breakup above an anomalously hot mantle may lead to the development of volcanic margins. Volcanic margins are characterized by (1) thick seaward dipping lava flow sequences, (2) central intrusive complexes associated with dyke swarms parallel to the coast, and (3) high seismic velocity bodies in the lower crust attributable to magma underplating. A conceptual model for volcanic margins development has recently been proposed based on onshore studies of the Greenland margins and the British Tertiary Igneous Province. It is proposed that the long-lived central intrusions are genetically linked to underlying persistent zones of mantle fusion. These localized melting domains (or soft spots), equivalent to small mantle diapirs, may locally soften the extending continental lithosphere. The low-viscosity diapirs would (1) localize tectonic strain and (2) feed the volcanic margin with magma. Thus such soft spots can control the along-strike magmatic and tectonic segmentation of volcanic margins. Recent geophysical investigations appear to show that the along-strike structure of volcanic passive margins is compatible with such a segmentation process. Here we present a set of scaled experiments designed to study how such localized rheological heterogeneities in the sub-Moho mantle may have a mechanical effect on continental breakup. Four-layer models were constructed using sand and silicone putties to represent the brittle and ductile layers of both crust and mantle. The soft spots are simulated by low-viscosity silicone putty emplaced within the brittle material. At the scale of the entire breakup zone, the

soft spots display an oceanic-type strength profile defining low-strength zones where continental breakup is initiated. The rift orientation and segmentation are strongly controlled by the distribution of the low-viscosity heterogeneities, rather than by the direction of regional extension. The experiments are compared with the geometry and segmentation of the onshore part of the Greenland volcanic margins. **INDEX TERMS:** 8109 Tectonophysics: Continental tectonics—extensional (0905); 8120 Tectonophysics: Dynamics of lithosphere and mantle—general; 8159 Tectonophysics: Rheology—crust and lithosphere; 8194 Tectonophysics: Instruments and techniques; **KEYWORDS:** analogue modeling, volcanic margin, Greenland, mantle diapir, soft point. **Citation:** Callot, J. P., L. Geoffroy, and J. P. Brun, Development of volcanic passive margins: Three-dimensional laboratory models, *Tectonics*, 21(6), 1052, doi:10.1029/2001TC901019, 2002.

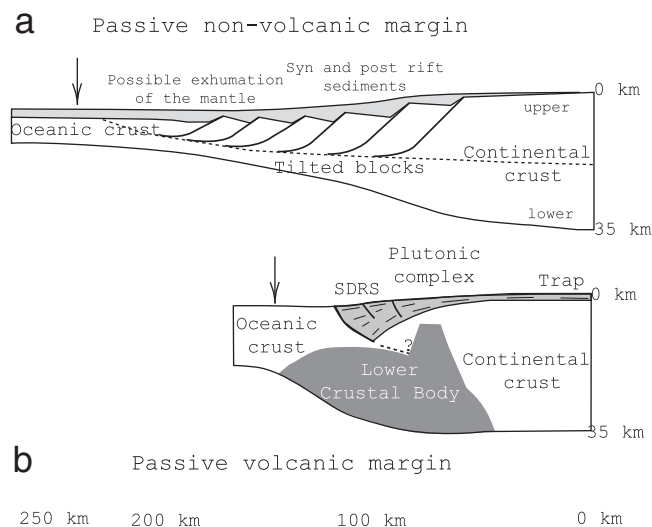
### 1. Introduction

[2] Volcanic passive margins (VPMs) belong to large igneous provinces (LIPs) that contain huge volumes of mafic extrusive and intrusive rocks emplaced in a short period of time [*Morgan*, 1981; *White and McKenzie*, 1989; *Richards et al.*, 1989; *Coffin and Eldholm*, 1994]. Such margins, associated with (1) an abnormally thick adjacent oceanic crust and (2) a hot spot track and tail, are related to lithospheric breakup over a mantle plume [*Holbrook and Kelemen*, 1993; *Eldholm et al.*, 1995].

[3] Since the early paper by *Morgan* [1971], several models have been invoked to describe the interaction between mantle plumes, continental breakup and LIPs and VPMs. *White and McKenzie* [1989] described lithospheric breakup as a passive pure-shear extension driven by plate boundary forces, in which the volume of erupted products depends solely on the potential temperature in the mantle beneath the rift zone. The preexisting rift zone here guide the ascent and subsequent melting of the hot plume. By contrast, impingement of a plume head below the lithosphere could result in a progressive thermal erosion of the lithosphere and in the development of a LIP, followed by lithospheric

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**Figure 1.** Simplified sketches of (a) nonvolcanic margin and (b) volcanic margin. Arrow indicates the continent-ocean boundary.

extension [Richards *et al.*, 1989; Griffiths and Campbell, 1991]. In such a case of active rifting, small-scale thermal convection generated within the plume head could rapidly thin the lithosphere from below and trigger localized magmatism after a more or less long period of incubation [Fleitout *et al.*, 1986; Griffiths and Campbell, 1991; Kent *et al.*, 1992]. On the contrary, models based on secondary convection have been described as an alternative to the plume hypothesis [Mutter *et al.*, 1988; Anderson, 1992; King and Anderson, 1995]. Steep lateral temperature gradients in the lithosphere could lead to localized secondary convection and melt generation from the thermal boundary layer of the cratonic lithosphere with an enriched geochemical signature, without the presence of a plume [Mutter *et al.*, 1988; Anderson, 1994; King and Anderson, 1995]. Plate breakup at major sutures or at craton edges, due to plate tectonic movements, could also lead to the rapid exhumation and melting of such a subcontinental mantle [Anderson, 1985, 1994]. Eventually, in a comprehensive review of the most important LIPs, Courtillot *et al.* [1999] propose a mixed active/passive scenario for rifting in the presence of a plume. This scenario emphasizes the causal relationship between the plume emplacement and the subsequent rifting, associated with the development of a LIP and, along the rift zone, of a VPM. In any case, the interaction between the mantle plume/abnormally hot mantle and continental lithosphere results both in (1) alteration and replacement of lithospheric material by asthenospheric plume-derived products, and (2) the emplacement of large volumes of melts within the lithosphere. Thus it considerably affects the thermal regime of the lithosphere and its mechanical properties as well [Saunders *et al.*, 1992; Frey *et al.*, 1998].

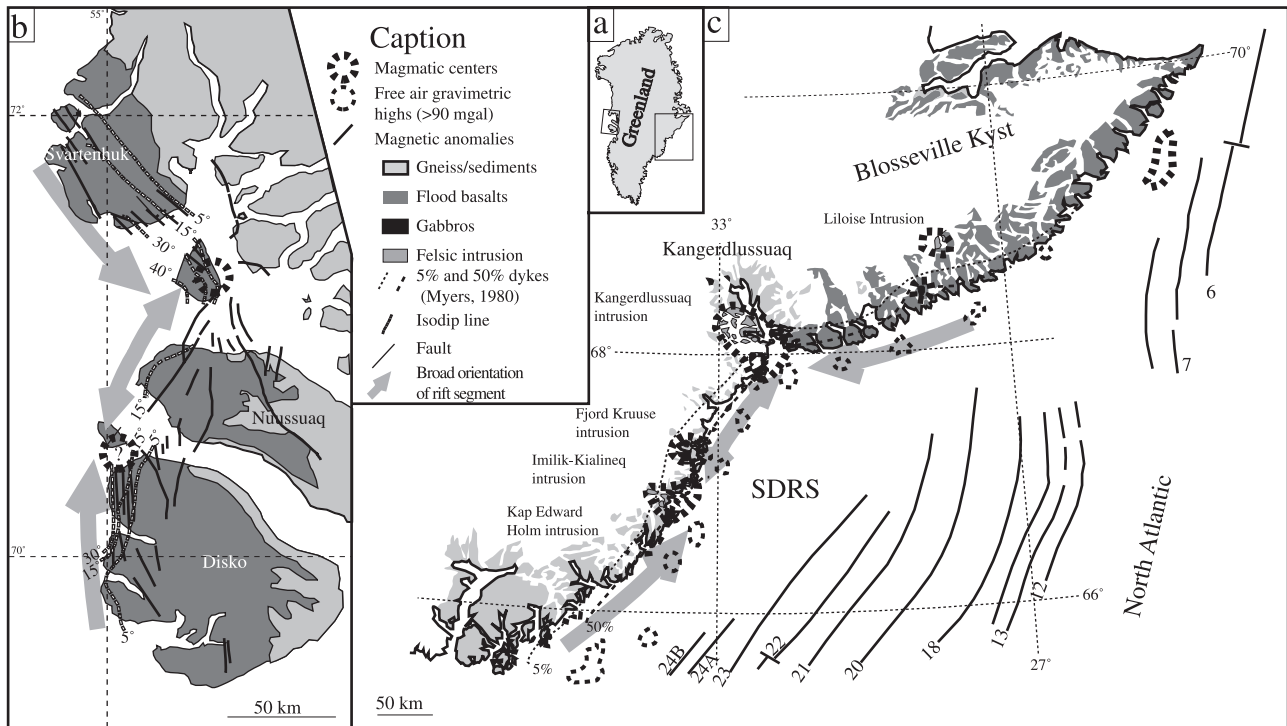
[4] At crustal level, VPMs record the interaction between continental lithosphere and the melting of an abnormally hot mantle [Kelemen and Holbrook, 1995]. They are characterized by (1) thick sequences of seaward dipping lava flows

erupted under subaerial conditions (seaward dipping reflector sequences, SDRS) [Hinz, 1981; Mutter, 1984; Eldholm *et al.*, 1995], (2) plutonic complexes associated with dyke swarms parallel to the coast [Myers, 1980; Barton and White, 1997], and (3) areas of high seismic velocity in the lower crust possibly related to magma underplating [White *et al.*, 1987; Eldholm, 1991; Eldholm and Grue, 1994; Kelemen and Holbrook, 1995; Frey *et al.*, 1998] (Figure 1). During continental breakup, VPMs undergo significant localized crustal thinning and stretching over a narrow area (roughly 100 km) [White, 1992; Barton and White, 1997; Eldholm *et al.*, 1995]. On the opposite to nonvolcanic passive margins where crustal thinning is marked on by large scale tilted blocks bounded by seaward dipping normal faults, VPMs are characterized by seaward dipping lava flows associated with arrays of presumably continentward dipping normal faults [e.g., Le Pichon and Sibuet, 1981; Nielsen, 1975; Tard *et al.*, 1991; Geoffroy *et al.*, 1998, 2001]. From a summary of the main features of the Greenland margins, we will illustrate the fact that, apart from the above described two-dimensional (2-D) structures, VPMs in Greenland also show essentially along-strike segmentation. From the available geological and geophysical data on VPMs and from extensive field work on the VPMs of Greenland, Geoffroy [2001] has recently proposed a conceptual model to explain the 3-D geometry and development of VPMs. This model emphasizes the causal link between persistent localized fusion zones in the asthenosphere that act as soft spots during extension of the lithosphere and the margin segmentation.

[5] In the present study, we examine the development of rifting in a multilayer brittle-ductile lithosphere model containing zones of low resistance that simulate areas of magma production and storage. We focus particularly on the geometry and segmentation of the rift zone in relation to the emplacement of low-viscosity bodies at depth. Callot *et al.* [2001a] have already studied the mechanical response to stretching of a continental lithosphere containing a low-viscosity 2-D layer which is emplaced at various levels within the rheological profile of the lithosphere. These results show that a thinning-type geometry, similar to that observed on volcanic margins at the scale of the lithosphere, may be due to interruption of the sub-Moho resistant mantle by the low-viscosity layer. Following these results, we now study the effect on rifting of localized soft spots that are still situated within the sub-Moho mantle. The present results are compared to the first-order geometry and segmentation of the well-constrained volcanic margins of Greenland.

## 2. Along-Strike Segmentation of North Atlantic Plume Related Margins

[6] The vast Thulean magmatic province has long been considered as resulting from the impact of the Icelandic plume beneath the Greenland European lithosphere during the Palaeocene [Upton, 1988; White and McKenzie, 1989]. The internal structures of VPMs related to the opening of the North Atlantic are seen at outcrop on both the east and west coasts of Greenland. They are associated offshore with



**Figure 2.** Volcanic margins of Greenland. (a) Localization of the studied areas; (b) simplified geological map of the West Greenland margin, with isotilt lines of the Tertiary lavas; (c) geological sketch map of the East Greenland margin. Arrow indicates strain gradient. Gravity highs above 90 mgal are likely attributable to plutonic complexes. Note the fairly regular spacing of the onshore plutonic complexes and offshore gravity highs.

thick “seaward dipping reflector sequences” [Hinz, 1981; Larsen and Jakobsdottir, 1988; Barton and White, 1997; Korenaga et al., 2000] and underplated magmatic bodies [Korenaga et al., 2000; Schlindwein and Jokat, 1999].

**2.1. West Greenland Volcanic Margin (Disko-Svartenhuk Area)**

[7] On the west coast of Greenland, Tertiary lavas crop out from the Svartenhuk Peninsula to Disko Island. At the continental end of the margin, these rocks consist of tabular flood basalts that are progressively down-flexured toward the sea (Figure 2b). Thick piles of breccias, tuffs, and lava flows fed by dykes were erupted during the Palaeocene [Larsen et al., 1992; Storey et al., 1998] and early Eocene [Storey et al., 1998; Geoffroy et al., 2001]. They are coeval with a long-lived central intrusive complex that was magmatically active for at least 20 Ma [Larsen et al., 1992] at the Island of Ubekjendt. Another possible complex proposed by Clarke and Pedersen [1976] is situated offshore NW of Disko. A striking feature of the West Greenland VPM is the existence of two subperpendicular stretching gradients [Geoffroy et al., 2001]. A seaward flexuring of the margin is associated with a seaward increase of horizontal stretching, accommodated by arrays of landward dipping faults. In addition, an along-strike increase of the flexuring toward the two igneous centers is well expressed by an increase in lava flow tilt

and dyke dilatation. Thus these two igneous centers localized strain during the breakup between the North American and Greenland plates [Geoffroy et al., 2001]. They also display an apparent segmentation of the volcanic rift, with a “zigzag”-type pattern and bending in their vicinity (Figure 2b).

**2.2. East Greenland Volcanic Margin (Kap Wandel-Scoresby Sund)**

[8] The East Greenland VPM includes the largest and thickest exposed flood basalts in the North Atlantic region [Eldholm and Grue, 1994] (Figure 2c). South of Kangerlussuaq, mafic plutonic complexes are at outcrop associated with “en echelon” dyke swarms cutting Precambrian gneisses due to deep erosion [Nielsen, 1975; Karson and Brooks, 1999]. The margin exhibits a seaward crustal flexure similar to that observed in West Greenland [Wager and Deer, 1938]. North of Kangerlussuaq, this flexure is well expressed by a seaward increase in tilt of the lava flows [Nielsen and Brooks, 1981] and, south of Kangerlussuaq, by an increase in seaward tilt of the earliest dykes [Karson and Brooks, 1999].

[9] A remarkable feature of the margin is the occurrence along strike of long-lived hypovolcanic complexes, associated with the dyke swarms, that mark out a distinct magmatic segmentation of the margin. Subcircular offshore gravimetric highs are attributable to similar complexes

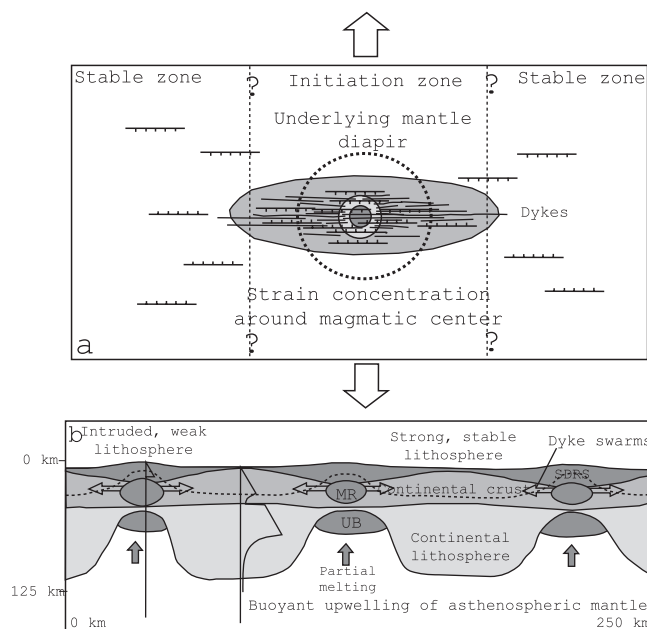
emplaced within the transitional crust [Barton and White, 1997; Korenaga et al., 2001]. Similar observations have been made on the Hebridean shelf [e.g., Bott and Tuson, 1973; Evans et al., 1989; Hitchen and Ritchie, 1993]. The magmatic activity covers two distinct periods, the older lasting from 57 to 55 Ma, and the younger from 50 to 47 Ma [Tegner et al., 1998]. Some areas show igneous activity up to 36–28 My [Noble et al., 1988; Tegner et al., 1998], indicating persistent localized zones of melting in the mantle beneath the complexes.

[10] Igneous units are imaged offshore that seem to display significant lateral variations in thickness of the lava flow pile as well as of the underplated body [Barton and White, 1997; Schlindwein and Jokat, 1999]. This could be related to lateral variations in the amount of mantle melting. At upper crustal levels, the magmatic segmentation is expressed onshore by “en echelon” patterns of dyke swarms (centered on intrusive complexes) [Myers, 1980; Karson and Brooks, 1999], a feature that is also detected by offshore aeromagnetic surveys [Larsen, 1978]. Moreover, the magma chambers located beneath the present-day intrusive complexes appear to represent the feeding centers for the Tertiary dyke swarms and overlying lava flows. This is illustrated by (1) the overall dyke swarm geometry [Klausen and Larsen, 2001], (2) the high degree of lava contamination by crustal material, as well as fractionation at low pressures [Saunders et al., 1998; Larsen and Saunders, 1998], and (3) dominantly horizontal magma flow within dykes [Callot et al., 2001b]. These plutonic complexes also appear to have localized strain, as indicated by the increased flexure in their vicinity and the curvature of the margin around the huge Kangerlussuaq complex [Myers, 1980; Nielsen and Brooks, 1981; Karson and Brooks, 1999; M. Broman Klausen, personal communication; our observations]. Thus the complexes define an along-strike tectonic segmentation of the margin (Figure 2c), rather similar to the one observed on the West Greenland margin.

### 3. A Conceptual Model of Volcanic Margin Segmentation and Development

[11] From the above, a major feature of the Greenland VPMs appears to be the localized nature of the melt sources in the mantle and their persistence through geological time. Melt production seems to be concentrated at points in the mantle which remained active for several million up to ten millions of years. These deep-seated melting points probably fed upper crustal magmatic chambers which in turn supplied both the central hypovolcanic intrusions and the coast-parallel regional dyke swarms through lateral injection [Callot et al., 2001b]. Thickness variations of the underplated bodies also suggest strong lateral inhomogeneities of melt production in the mantle, which could be correlated with the observed spacing of their upper crustal expression, i.e., the distance between central complexes.

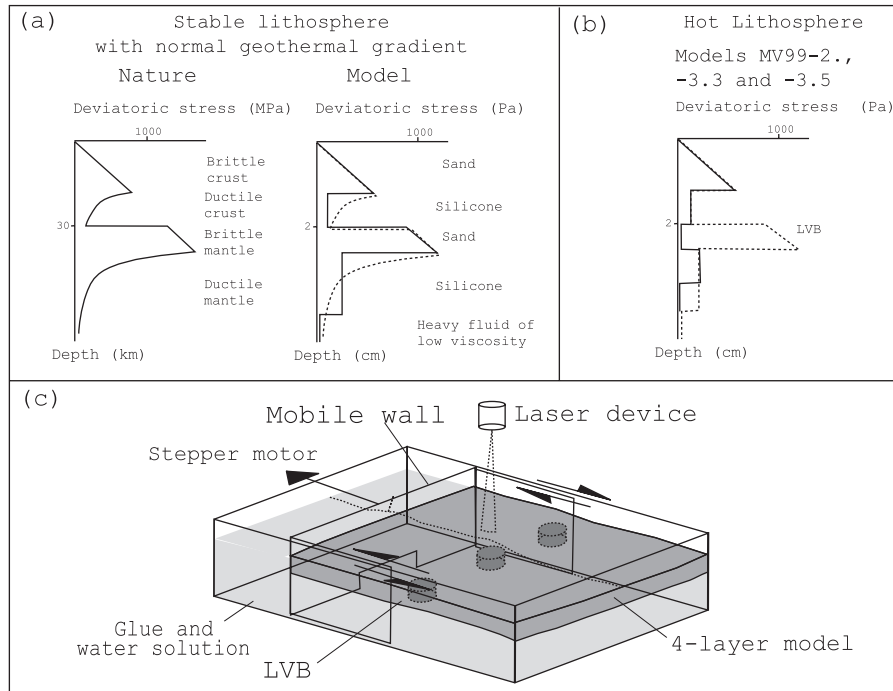
[12] The localized nature of melt production and ascent at volcanic margins is also suggested by other studies of VPMs. Behn and Lin [2000] suggest that the along-strike variation of the coast-parallel isostatic anomaly of the U.S.



**Figure 3.** Conceptual model of volcanic margin segmentation. (a) Plan view of an initiation segment bounded by stable zones; (b) along-strike vertical cross section; both modified after Geoffroy [2001] (reprinted from *Marine Petroleum Geology*, with permission from Elsevier Science). MR: magma reservoir; UB: underplated body; SDRS: seaward dipping reflectors.

East Coast margin may be related to variations in the amount of underplated material. A similar pattern of offshore free-air gravity anomaly is observed on the Namibian margin [Light et al., 1992; Bauer et al., 2000]. These lateral variations exhibit a spacing of the order of 100 km, close to the segmentation of slow-spreading ridges and close to the apparent spacing of the onshore central complexes in East Greenland (Figure 2). Some well-defined rounded gravimetric highs in the crust are directly associated with igneous complexes characterized by high seismic velocities throughout the crust [Bauer et al., 2000].

[13] On the basis of the previous observations and extensive fieldwork, Geoffroy [2001] proposed that small-scale thermally driven convection within the plume head could lead to the development of long-lived melting spots. The central complexes and underplated bodies associated with these melting zones would localize both magmatic accretion and distension within the crust (Figure 3a). Geoffroy [1998] suggests that the development of a persistent asthenospheric fusion zone in the continental mantle, as well as the associated melt reservoirs up to the Moho and in the crust, should locally weaken the stable continental lithosphere and focus extensional strain, thus initiating a local rift zone (Figure 3). The association of a localized mantle fusion zone and crustal magma reservoirs could be responsible for along-strike variations in strain during the development of a volcanic margin. In this way, it may define the basic segmentation of continental breakup at lithospheric scale (Figure 3b). This model may be compared



**Figure 4.** Strength profiles. (a) In nature and in a four-layer lithosphere containing no LVB; (b) for illustrated experiment with a LVB; (c) experimental apparatus.

to *Courtillot's* [1982] concept of rift initiation and growth. In *Courtillot's* model, rifts nucleate at favorably oriented weak zones in response to plate-driving mechanisms. The rate of rift propagation slows down as it reaches more resistant zones. Differential extension leads to the formation of “margin plateaus” and defines an along-strike segmentation of the passive margin [Dunbar and Sawyer, 1996]. In *Geoffroy's* [2001] model, the development of several localized zones of fusion results in an along-strike segmentation of the margin through mechanical interaction between the growing rift segments (Figure 3b).

## 4. Laboratory Experiments

### 4.1. Experimental Procedure

[14] The rifting of a rheologically layered continental lithosphere has been studied by numerical modeling [Dunbar and Sawyer, 1989, 1996] and analogue experiments [Brun and Beslier, 1996; Brun, 1999; Callot et al., 2001a]. Until now, numerical models have failed to simulate the brittle behavior of crustal and sub-Moho mantle layers. On the other hand, analogue experiments are unable to reproduce the thermal dependence of rheological parameters, but have proven successful in studying the bulk and internal mechanical instabilities of an extending lithosphere [e.g., Brun, 1999].

[15] The models described below are scaled to represent the passive evolution of a continental lithosphere subjected to regional extension. A continental lithosphere subjected to extension at medium strain rates ( $10^{-14}$  to  $10^{-15}$  s<sup>-1</sup>), with a normal thickness of crust (35 km) and a stable continental

geotherm (30°C/km), can be represented by four-layer type strength profiles (Figure 4a) [Dunbar and Sawyer, 1989; Buck, 1991; Brun and Beslier, 1996; Brun, 1999]. The multilayer lithosphere is modeled using sand and silicone to represent respectively the brittle and ductile layer of both crust and mantle. The lithosphere floats over a low-viscosity higher density material which represents the asthenosphere. The thickness, viscosity, density, stresses and strain rate are appropriately scaled (see scaling principles given by Davy and Cobbold [1991] and Brun [1999]). Scaling and physical parameters are summarized in Table 1 and Figures 4a and 4b for the different models presented below. The model is set up in a rectangular tank with one mobile wall connected to a computer controlled stepping motor.

[16] The resistant brittle layers (brittle crust, Bc, and brittle mantle, Bm) are simulated by Fontainebleau sand sieved at 400 μm. Sand exhibits a perfect Mohr-Coulomb behavior with a mean frictional coefficient ( $\phi$ ) of 0.58 and a density ranging from 1400 to 1750 kg m<sup>-3</sup>, depending on whether it is poured or sifted [Krantz, 1991]. The ductile lower crust (Dc) and the lower part of the lithospheric mantle (Dm) are modeled using silicone putties. These silicones exhibit viscous Newtonian behavior with a viscosity ( $\mu$ ) ranging from  $1.3 \times 10^4$  to  $2.5 \times 10^4$  Pa s at 20°C and a density of 1250 kg m<sup>-3</sup>. The lithospheric part of the model floats above a low-viscosity layer of higher density material representing the asthenospheric mantle, which allows free isostatic compensation. In some models, the material is a water-based gel with a nearly perfect Newtonian rheology composed of commercial glue and sodium polytungstate to control viscosity and density, respectively (density: 1450 kg m<sup>-3</sup>; effective viscosity: 10 to 100 Pa s). This material

**Table 1.** Experimental Parameters for Models Shown in Figures 3–10

Model	Initial Model Dimension (Length × Width), cm	Initial Layer Thickness of Brittle/Ductile Crust and Mantle <sup>a</sup> , cm	Physical Properties of Silicone Putty <sup>b</sup>	Deformation Parameters <sup>c</sup>
MV 99-2.1	25 × 22	Bc/Dc/Bm/Dm $\frac{1.5/1.2//0.9/1.4}{5}$	–	v = 1; d = 300; e = 5
MV 99-3-2 MV 99-3.3	50 × 46	Bc/Dc/Bm/Dm $\frac{1.3/1//1/1.3}{4.6}$	Dc: D = 1.2; $\mu = 1.8 \times 10^4$ Dm: D = 1.4; $\mu = 3.8 \times 10^4$ LVB: D = 1.2; $\mu = 3 \times 10^3$	v = 1; d = 300; e = 5 (MV 99-3-2); e = 1 (MV 99-3.3)
MV 99-3-5 MV 99-3.6	50 × 46	Bc/Dc/Bm/Dm $\frac{1.2/1//1/1.3}{4.5}$	–	v = 1; d = 240; e = 4 (MV 99-3-5); e = 1.2 (MV 99-3.6)

<sup>a</sup> Abbreviations are as follows: Bc, brittle crust; Dc, ductile crust; Bm, brittle mantle; Dm, ductile mantle.

<sup>b</sup> Abbreviations are as follows: D, density;  $\mu$ , mantle; LVB, low-viscosity bodies.

<sup>c</sup> Abbreviations are as follows: v, mobile wall velocity (in cm hr<sup>-1</sup>); d, experimental duration (min); e, extension length (cm).

allows the model to be frozen and cut. In some other types of model, the asthenosphere is modeled using a glucose syrup solution with a slightly non-Newtonian behavior (density: 1450 kg m<sup>-3</sup>; effective viscosity: 50 to 300 Pa s), but in this case the models cannot be frozen and cut. Brittle layers are built up with sand of various colors to pick out the shear zone geometry within the brittle parts. Silicone layers contain passive vertical markers whose deformation illustrates the bulk flow pattern within the layer.

[17] We introduced local heterogeneities made of silicone putty of low viscosity to model the soft spots. These low-viscosity bodies (LVB) show a viscosity contrast of at least ~1 order of magnitude with respect to the surrounding parts of the lithosphere (Table 1). Previous experiments carried out by *Callot et al.* [2001a] simulated the cross-sectional geometry and rheological layering of volcanic margins. They showed that the behavior of LVBs is similar to the breakup of a volcanic margin, leading to a rapid and localized breakup if the LVB crosscuts the high-strength sub-Moho mantle. Here, we focus on the role played by the local emplacement of more or less cylindrical structures composed of low-strength mantle fusion zones through the continental lithosphere. On the basis of *Callot et al.* [2001a] results, such structures are emplaced within the mantle lithosphere up the Moho. The apparent along-strike variations of volcanic margins in the North Atlantic display a spatial periodicity for first-order structure, i.e., the distance between major crustal plutonic complexes is roughly 70 to 100 km [*Karson and Brooks*, 1999; *Geoffroy et al.*, 2001]. We therefore model the LVB as a cylindrical body 4 cm in diameter, located within the sub-Moho mantle and separated by 4 cm of unaltered lithosphere.

[18] Several drawbacks arise from the choice of analogue material, as well as from the laboratory constraints and the boundary conditions. Displacements are applied through velocity discontinuities located at the middle of the model walls. Such boundary conditions are not very restrictive since the width-thickness ratio of the models is high. This is illustrated by the limited extent of oblique structures near the model walls. A major drawback concerns the scaling of temperature and rheological properties. As there is no heat

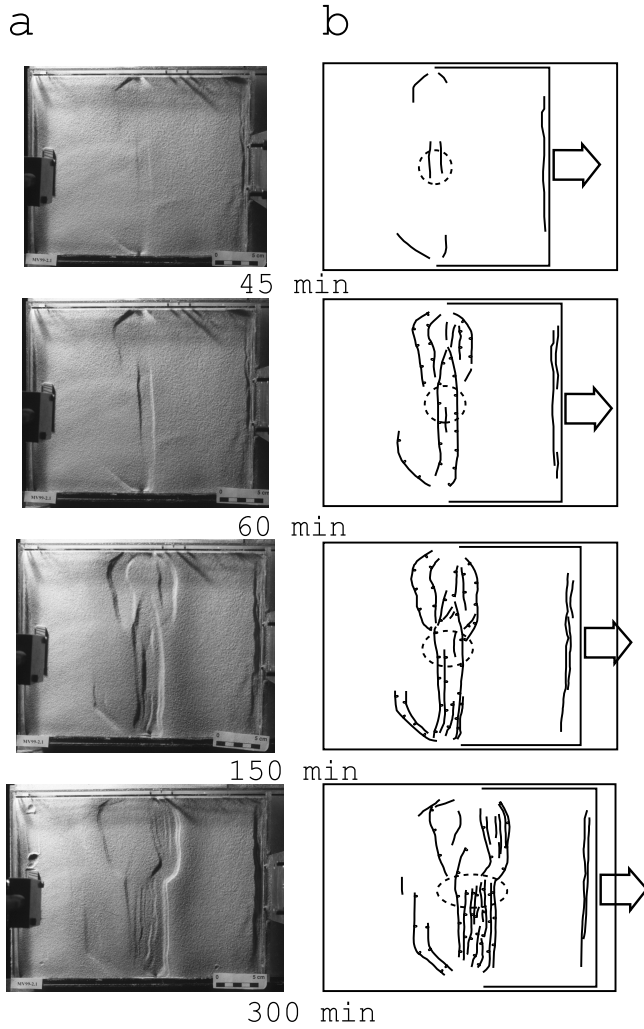
diffusion and no temperature dependence of rheological properties in analogue materials, no account can be taken of thermal recovery which would restore the lithosphere thickness and equilibrium. Therefore the models cannot simulate strength variations induced by temperature changes during deformation.

#### 4.2. Effect on the Deformation Pattern of a Single Localized Soft Spot

[19] In Figure 5, we present the surface deformation of a model containing a single LVB. Serial cross sections are shown in Figure 6. Deformation in the upper brittle layer does not appear at the onset of extension. The first graben initiates after 40 min, i.e., after 0.66 cm of finite displacement, at the top of the LVB (Figure 5a). At least two grabens initiate on both sides of the model at the velocity discontinuities between the mobile wall of the piston and the stable sidewall of the box. The central graben rapidly propagates toward the lateral walls of the box. After ~1 cm of finite displacement, the deformation pattern at the surface shows three distinct domains of crustal thinning.

[20] The first domain consists of a central narrow zone of stretching above the LVB, where the stretching is accommodated by a single graben (Figures 5 and 6). For an applied finite displacement approaching the thickness of the lithosphere, the width of the deformed zone does not exceed the thickness of the lithosphere. During the entire deformation, stretching is strongly localized above the LVB. Due to contraction and leaking of asthenosphere material during freezing of the model after the experiment, the central part of the model (Figure 6b) undergoes a slight subsidence altering the geometry of the structures above the LVB. Nevertheless, the style of breakup is significantly different from classic models of necking in a four-layer lithosphere [*Brun*, 1999]. A remarkable feature is the lack of large-scale tilted blocks in the brittle crust. The lithospheric breakup is similar to that described by *Callot et al.* [2001a] for 2-D models.

[21] On the contrary, along-strike on both sides of the LVB, the thinned domain is twice as large as above the LVB and



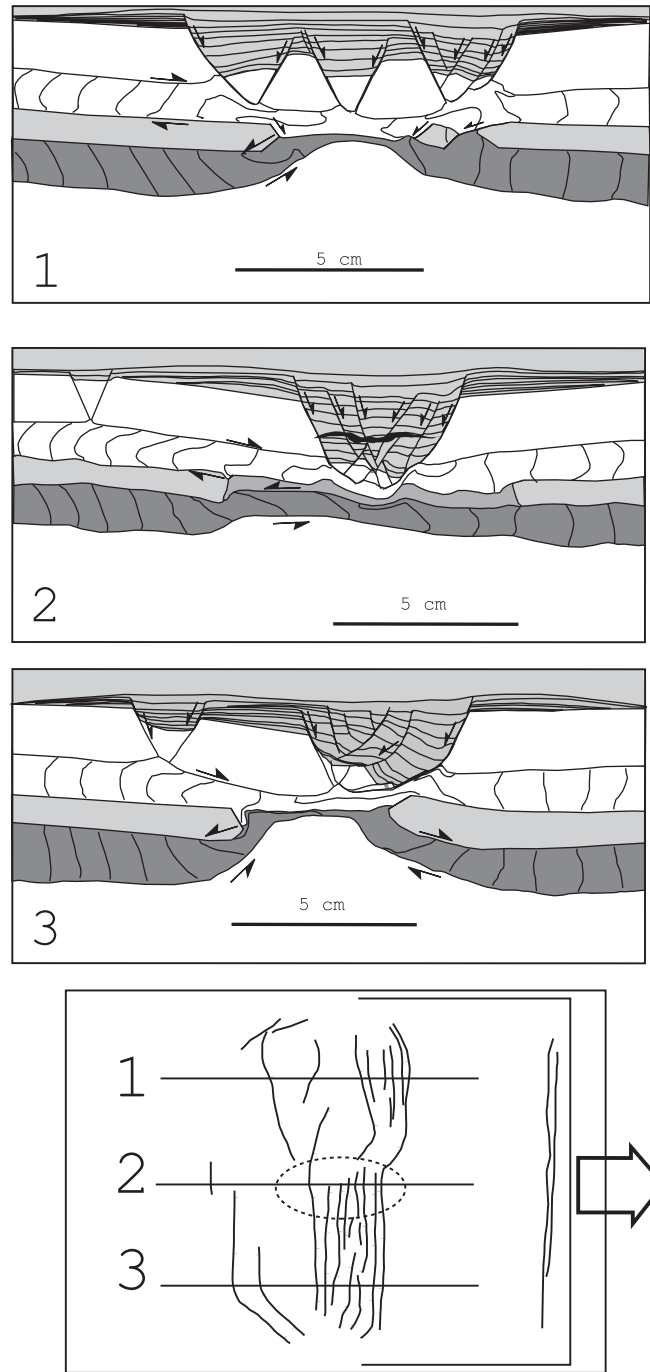
**Figure 5.** Experimental results. (a) Plan view of model MV99-2.1 at different stages of evaluation; (b) interpretative sketches; note the absence of transfer structures between adjacent rift segments.

consists of a succession of horsts and grabens (Figures 5 and 6). On the lower side of the model surface views, a rollover structure develops due to strike-slip shear along the side wall of the box (Figure 6, below). The style of deformation is fairly similar to classic models for the breakup of continental lithosphere [Brun and Beslier, 1996]. Necking is accommodated through development of lithospheric-scale conjugate shear zones. Since no transfer fault zone is observed between the three domains, an important reorganization of extensional structures occurs in the vicinity of the LVB. Instead, we observe a progressive curvature toward the LVB of the faults bounding the major grabens, a pattern particularly well expressed on the upper side of the model surface (Figure 5).

**4.3. Interaction Between Soft Spots and Effects on Rift Geometry**

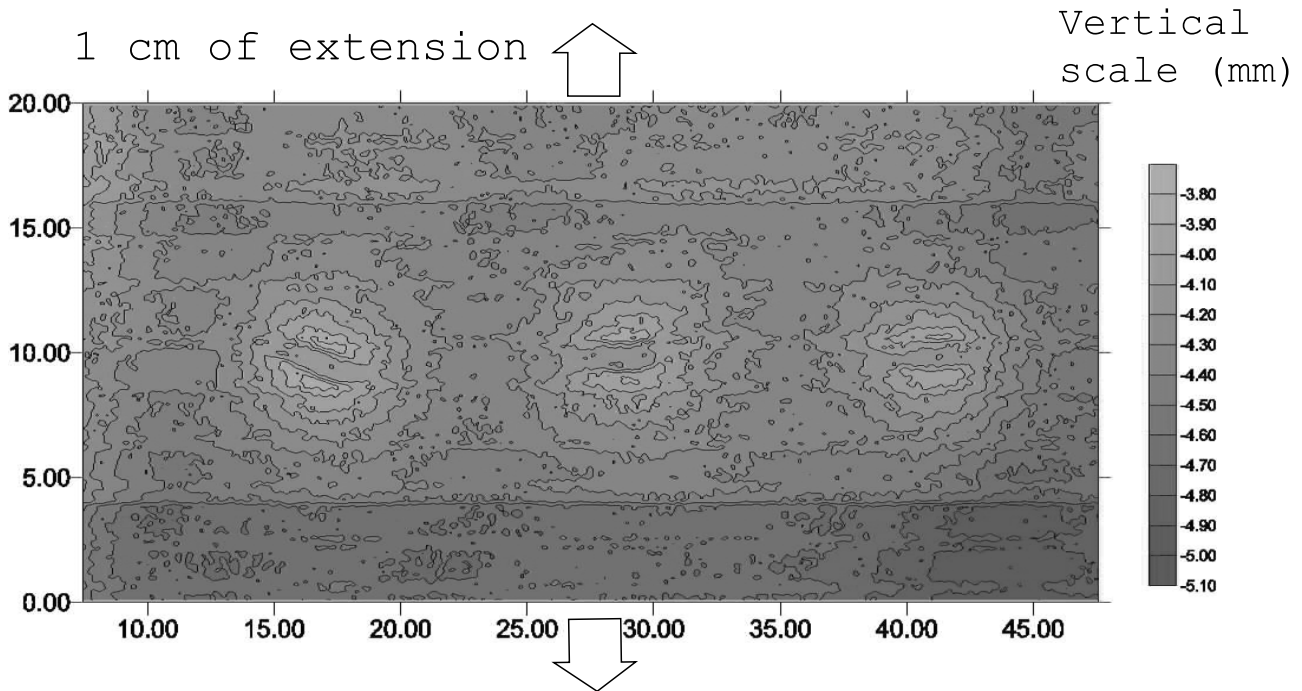
[22] In Figures 7–9, we show some experiments designed to test the combined effects of several localized

low-viscosity bodies on rift initiation and segmentation. In the two first models, three LVBs are aligned perpendicular to the stretching direction in order to study the margin segmentation (Figures 7 and 8). In another model (Figure 9), three LVBs define a V-shaped geometry whose axis of symmetry is parallel to the stretching direction. In the experiments designed to incorporate several LVBs, the large width of the model makes it technically difficult to cut



**Figure 6.** Serial cross sections of model MV99-2.1 (see Figure 5).





**Figure 7.** Early stage of stretching of model MV99-3.2 (similar to MV99-3.3, Figure 8), containing three aligned soft spots, laser topographic map. A single narrow graben initiates above each LVB.

vertical sections. For all aspects requiring vertical cross sections, we therefore rely on the results obtained from smaller models with a single LVB (model MV99-2.1; Figures 5 and 6) as well as the results of *Callot et al.* [2001a].

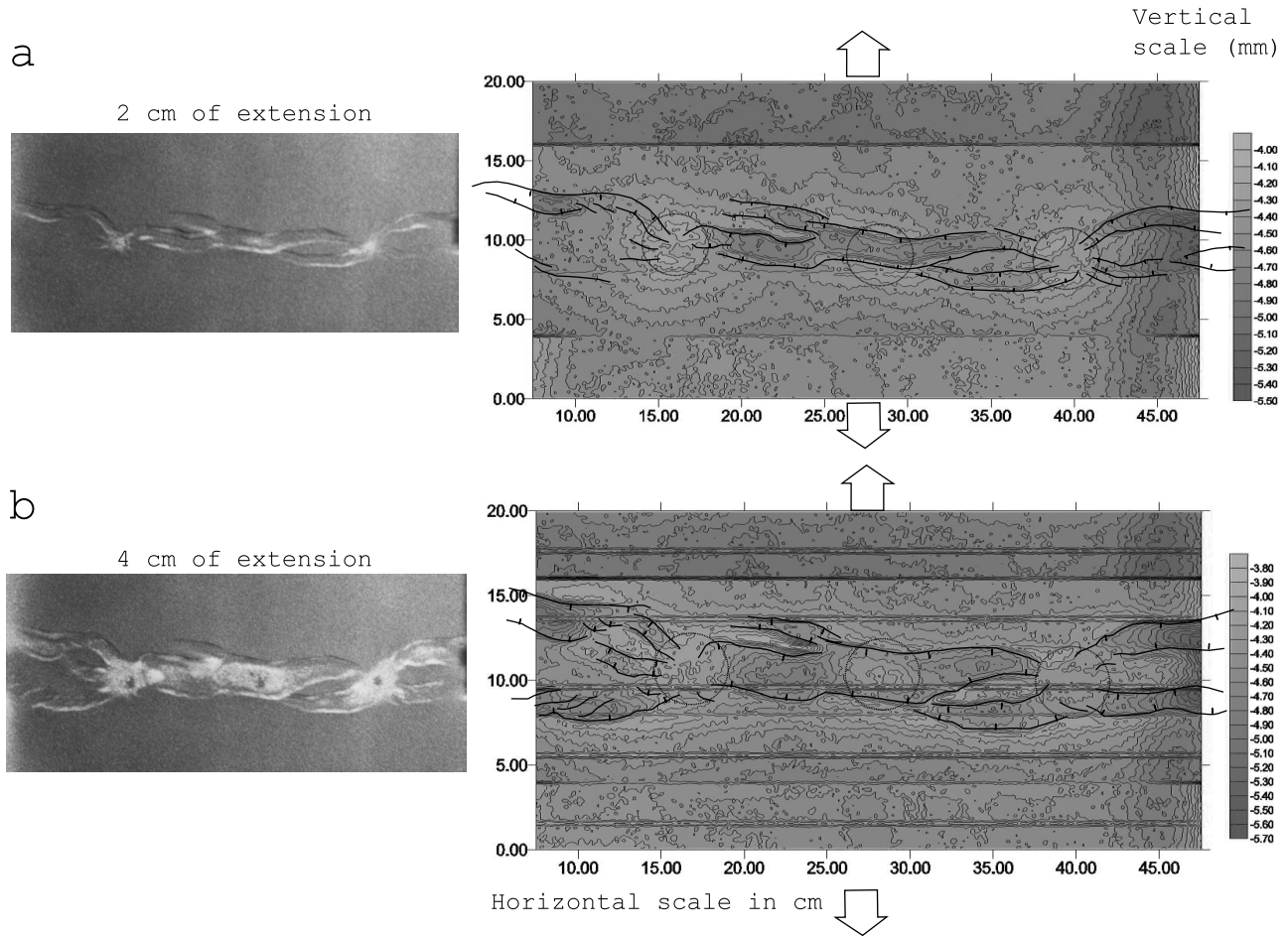
[23] In case of linear arrays of LVBs, a single graben initiated on top of each LVB. This phenomena was particularly well expressed in experiment MV99-3.2 (Figure 7), but later on during the experiment, the model failed near the mobile wall. Stretching was then entirely accommodated along the mobile wall. During experiment MV99-3.3, which is similar to experiment MV99-3.2, a single graben is initiated solely above the central LVB (Figure 8a). Above the two other LVBs, stretching is accommodated by homogeneous thinning of the brittle crust, due to imperfect local model building. Twin grabens are then initiated at velocity discontinuities and later between the LVBs. Soft spots act as zones of initiation where stretching is able to nucleate. Grabens later propagate toward the stable parts of the model with four layers. LVBs delineate narrow rift segments consisting of single grabens that separate wide rift segments made up of twin grabens. The exhumation of mantle layers is quickly achieved within the narrow rift segments after only 2.5 cm of finite displacement. Isolated horsts within the wide rift segments may suffer rotation due to propagation and overlapping of the segments (compare Figures 8a and 8b). At the end of the experiment, the width of narrow rift segments is comparable to the applied displacement. By contrast, the width of the wide rift segment is at least 50% more than the applied displacement. A remarkable feature of the model is the focusing of stretching above the LVBs. The transition from single to twin rift segments is accommo-

dated by a progressive curvature and reorientation of the grabens toward the LVB rather than by the formation of transfer fault zones. At the end of the experiment, the breakup zone is almost continuous and displays three well-defined rift segments.

[24] Volcanic margins currently display strong curvature and reorientation that are possibly related to the presence of localized mantle fusion zones and their associated magmatic plumbing (Figures 2 and 3). Because of this, the experiment in Figure 9 was designed to model the effect of nonaligned LVBs on the passive stretching of a stable continental lithosphere. As in previous experiments, rift zones initiate above the LVBs as well as at the velocity discontinuities along the model edges. Then, they rapidly propagate across the stable parts toward the adjacent soft spots. After 3 cm of finite displacement, the rift zones are connected up in an almost continuous structure that is strongly oblique to the imposed direction of displacement in the neighborhood of the soft spots. Fault curvature is first accommodated by “en echelon” sets of normal faults between the soft spots (Figure 9a). The connection between these faults is almost achieved at the end of experiment after 4 cm of finite displacement (Figure 9b). In this case, a “continental block” begins to come apart due to the inception and propagation of an oblique normal fault zone, i.e., a normal fault between two of the LVBs, with a dextral strike-slip component.

#### 4.4. Comparison With the Greenland Margins

[25] The experiment shown on Figure 9 exhibits an overall rift geometry fairly similar to the structural map of

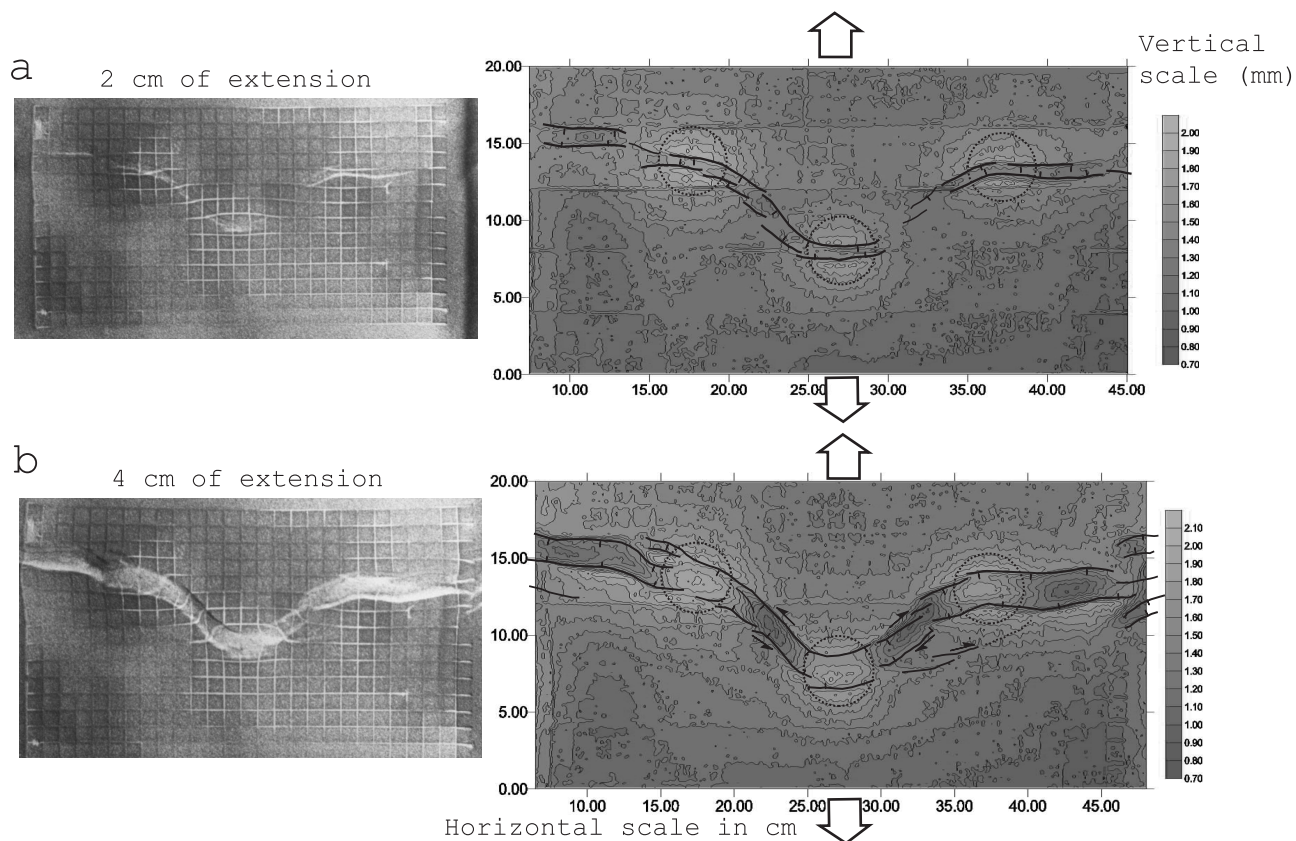


**Figure 8.** Experimental results for model MV99-3.3 containing three aligned soft spots at different stages of the deformation, together with laser topographic maps.

the East Greenland VPM. This supports our hypothesis that soft spots in the model correspond to major zones of magmatic activity along the Greenland margin. From north to south, these comprise: the Liloise complex, the Kangerlussuaq-Skaergaard-Kap-Edward complex and the Kap-Gustav-Imilik-Qialineq complex (Figures 2 and 10). The initial geometry of extension at chron 24, i.e., the geometry of the coastal flexure, seems fairly comparable to the present day reentrant of the margin. However, it should be noted that the model was not constructed specifically to represent the east Greenland VPM. Our model reproduces very well the apparent along-strike segmentation of the East Greenland margin [Myers, 1980; Karson and Brooks, 1999], although the assumed cause of segmentation is different. This segmentation is marked by strain localization, well illustrated by an increase in dyke tilting in and around the central complexes, and separated by areas of low strain intensity. This is comparable to the change in rift width in the models with aligned LVBs. Such a feature is not observed for staggered LVBs (see below). Finally, we may tentatively compare the separation of the Jan Mayen microcontinent from the Greenland plate with the late stage of experiment 3.5 (Figure 9b). The fault interaction between two LVBs

in this experiment gives rise to a secondary rift zone that progressively leads to the separation of a 8-cm-long continental block, a length corresponding to roughly 160 km in nature. Such a length is of the same order of magnitude as the Jan Mayen block (~250 km).

[26] A similar comparison may be drawn between the West Greenland margin and model MV99-3.3. The West Greenland margin bends around two crustal plutonic complexes, located at Ubekjend Ejland and northwest of Disko, with a spacing of the order of 100 km (Figures 2b and 11). However, the stretching direction is different here because the West Greenland margin was subjected to a regional NE-SW trending distension during rifting [Geoffroy et al., 2001]. The northern and southern rift segments of Svartenhuk and Disko seem to be controlled by this regional distension away from the soft spots. These rift segments should correspond to the external part of the rift zone of model MV99-3.5. As such, this should express a dynamic similarity between model and nature. This is because the stretching direction with respect to the mean rift trend outside the soft spot perturbations is comparable to the synmagmatic stretching direction in nature. However, the rift segment between the two soft spots of Hareoen and Ubekjend Ejland is much more oblique to the stretching



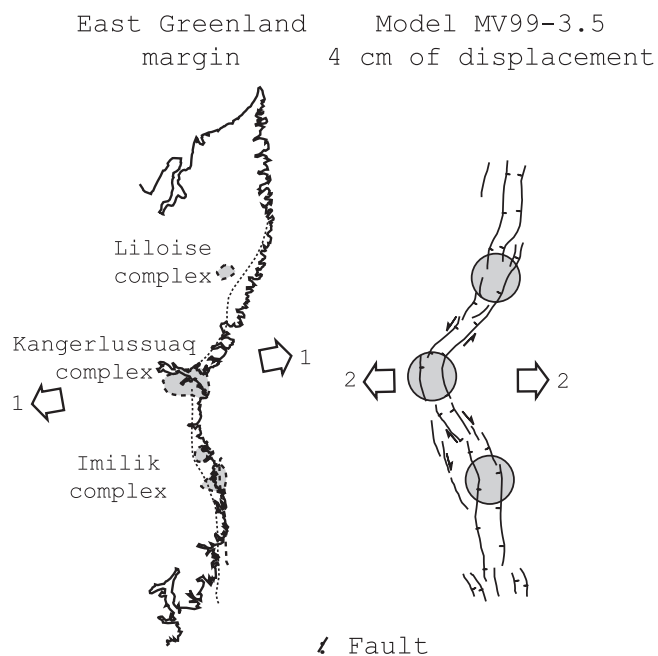
**Figure 9.** Experimental results for model MV99-3.5 containing three nonaligned soft spots at different stages of the deformation, together with laser topographic maps.

direction than in our model (Figures 2b and 11). Despite this discrepancy, the comparison is reasonable since model MV99-3.5 shows that the rift direction is more controlled by the soft spot distribution than by the direction of stretching. Moreover, the sinistral strike-slip component of strain along the extensional Nussuaq “central segment” [Geoffroy *et al.*, 2001] is comparable to the observed sinistral sense of shear in model MV99-3.5.

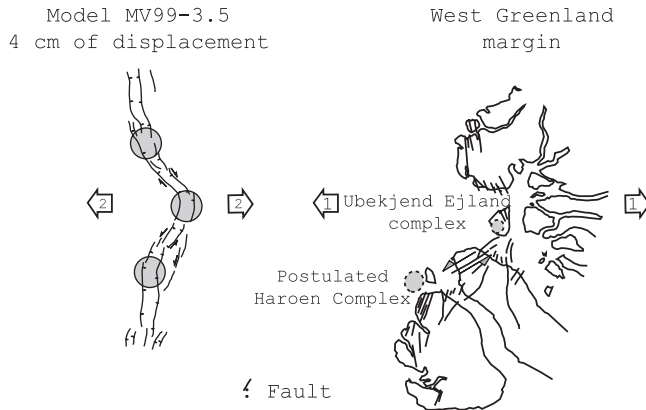
## 5. Discussion

### 5.1. Patterns of Rift Initiation

[27] Lithosphere models containing LVBs within the sub-Moho mantle exhibit a two-layer type rheological profile with the highest strength material at the base of the brittle crust. The surrounding parts of models consist of four layers, with the highest strength occurring below the Moho. In case of a two-layer type profile, stretching is initiated at a single velocity discontinuity (VD) at the interface between upper brittle and lower ductile crust, thus giving rise to a single symmetrical graben [Allemand *et al.*, 1989; Allemand and Brun, 1991; Brun, 1999]. On the contrary, within the parts of the model with four layer, stretching is accommodated by two conjugate normal-sense shear zones initiated at the two VDs at the base of the brittle crust [Brun and Beslier, 1996; Michon and Merle,



**Figure 10.** Comparison of model MV99-3.5 (Figure 9) with the East Greenland margin. Onshore plutonic complexes are related to their corresponding mantle diapirs, i.e., soft spots.



**Figure 11.** Comparison of model MV99-3.5 (Figure 9) with the West Greenland margin (see text for details).

2000]. Twin grabens are therefore developed, giving a wider deformation zone in the brittle crust. In the case of nonaligned LVBs, a unique graben develops even within the four-layer parts of the model. This particular deformation pattern is due to the obliquity between the stretching direction and the rift segment developing between the LVBs. From a mechanical point of view, this corresponds to an asymmetrical VD at the base of the lithospheric part of the model, giving a unique VD at the base of the brittle crust [Allemand *et al.*, 1989]. This boundary condition leads to the inception of a single symmetrical graben within the brittle crust.

## 5.2. Narrow Rifting, Wide Rifting, and Ultranecking

[28] Extension of the continental lithosphere can occur according to two different mechanisms [Brun, 1999]: (1) narrow rifting, leading to passive margin formation and continental breakup [e.g., England, 1983], and (2) wide rifting involving core complexes such as in the Basin and Range [e.g., Hamilton, 1987]. However, Buck [1991] identifies core complex formation as a separate type of mechanism.

[29] Necking of the whole lithosphere leads to the formation of passive margins and occurs in resistant continental lithosphere whose mechanical behavior is dominated by a high-strength sub-Moho mantle [Dunbar and Sawyer, 1989; Buck, 1991; Brun and Beslier, 1996]. The continental breakup produces symmetrical necking over a distance of around 200–400 km. At the lithospheric scale, deformation corresponds to pure shear even if simple shear can develop locally within the deformed domain [Govers and Wortel, 1995; Brun and Beslier, 1996]. The width of the necking zone is reduced either by decreasing the coupling between brittle and ductile layers, i.e., increasing the strain rate [Kuznir and Park, 1987; Buck, 1991; Brun, 1999], or by local weakening of the mantle by increasing the crustal thickness [Dunbar and Sawyer, 1989]. As strain rates are likely to be high in VPMS [Frey *et al.*, 1998; Geoffroy *et al.*, 2001], an extremely narrow necking of the lithosphere may be expected for VPMS. Moreover, a high-strength lithospheric mantle would have been thinned by heat conduction above the locally uprising asthenosphere. Thus the strength

profile of the lithosphere above a LVB displays a maximum at the base of the upper crust, which is due to the interruption of the sub-Moho high-strength layer. Our experiments, as well as previous results [Brun and Beslier, 1996; Michon and Merle, 2000; Callot *et al.*, 2001a], show that a local interruption of the sub-Moho high-strength material strongly controls the localization of stretching and will lead to lithosphere necking over a narrower zone than in a four-layer type lithosphere. In other terms, this corresponds to ultranecking, a pattern that is consistent with the overall thinning profile of VPMS.

[30] The strength profile above an LVB, which consists of a single brittle layer lying above the ductile lithosphere, can be compared with the strength profile of either: (1) an old cratonic lithosphere equilibrated during Precambrian times, (2) a hot continental lithosphere, or (3) a thickened oceanic lithosphere [Tapponier and Francheteau, 1978; Buck, 1991; Lin and Parmentier, 1989; Brun, 1999]. In a thickened and thermally relaxed continental lithosphere, gravity-driven spreading of the crust is the most likely response to buoyancy forces, allowing wide rifting as seen in the Aegean Sea or the Basin and Range [Buck, 1991; Brun, 1999]. In a normal thickness continental lithosphere undergoing an increase in geothermal gradient, the coupling between crustal and mantle layers decreases along with the strength of the sub-Moho, resulting in a general widening of the zone of necking [Dunbar and Sawyer, 1989; Buck, 1991]. In comparison, oceanic rifts are rather narrow, the stretching of oceanic lithosphere being more generally described in terms of steady state necking [Tapponier and Francheteau, 1978; Lin and Parmentier, 1989] due to its low viscosity and density as well as the contrast between crustal and mantle layers. VPMS should correspond to a narrow necking zone, comparable to the necking of a thick oceanic lithosphere.

## 5.3. Margin Segmentation

[31] Our experiments show complex patterns of rift initiation and segmentation. Narrow grabens nucleate above the LVBs at early stages of the deformation history and then propagate to the surrounding lithosphere. In case of LVBs aligned perpendicular to the stretching direction, twin grabens develop between the LVBs. Rifting in models displays a progressive along-strike segmentation of the margin that is comparable, at a different scale, to Courtillot's [1982] model. Above LVBs, the two-layered nature of the lithosphere gives rise to a weak zone where breakup can rapidly occur. On the contrary, the four-layer lithosphere between LVBs suffers a more diffuse deformation for a longer period before breakup, becoming a so-called "locked zone" [Courtillot, 1982; Dunbar and Sawyer, 1996]. Each soft spot defines the center of an accretionary domain that corresponds to a margin segment [Geoffroy, 2001]. Stretching is concentrated within a narrow rift zone, in the vicinity of the soft spot, and widens away from the soft spots. The transition from one rift segment to another is rather diffuse and is never accommodated by sharp boundaries such as transform faults. The segmentation of volcanic margins in our models is thus similar to the succession of narrow margins and margin plateaus [Dunbar and Sawyer, 1996], with a typical spacing between the magmatic com-

plexes of the order of 100 km. At a larger scale, the entire volcanic margin probably defines the zone of initiation of the lithosphere breakup. This passes laterally to margin plateaus with more classical patterns of crustal stretching like those seen on the Newfoundland, North Biscay, or Galician margins, although these examples are diachronous in nature [Keen *et al.*, 1989; Boillot *et al.*, 1989; Brun and Beslier, 1996; Dunbar and Sawyer, 1996]. Marginal plateaus are not observed around the Greenland plate because it was bounded by spreading ridges that were active during the early Paleocene. However, the South Atlantic volcanic margins are bounded to the south by the Falklands continental shelf, and the U.S. East Coast margin is bounded by the Newfoundland and Black Spur plateaus.

## 6. Conclusions

[32] Analogue modeling of a stable continental lithosphere containing low-viscosity bodies leads to new results on the development of volcanic margins and their segmentation. On a broad lithosphere scale, the development of a narrow deformation zone leading to fast and localized breakup of the lithosphere. This is linked to the local replacement of the sub-Moho brittle layer by a low-viscosity body which represents the emplacement of a small-scale mantle diapir corresponding to a long-lived localized fusion zone. During extension, local low-viscosity bodies induce an along-strike segmentation of the rifted domain. Rift segments initiate above the low-viscosity bodies and propagate toward adjacent segments. Breakup is strongly delayed within the four-layer part of the model, which acts as a more resistant zone developing small-scale margin plateaus. In the case of nonaligned viscosity heterogeneities,

the rift orientation seems to be controlled principally by the position of these heterogeneities rather than by the extension direction, in a way directly comparable to the Greenland volcanic margins. Our models emphasize the crucial role played within the lithosphere by the emplacement of small-scale intrusions of hot mantle material from the uprising asthenosphere. It is noteworthy that large igneous provinces preferentially form at craton boundaries [e.g., Anderson, 1994; King and Anderson, 1995], and this may be related to possible edge effects due to a strong variation in lithosphere thickness. Such a lateral contrast of temperature may channel the ascent of plume material [Thompson and Gibson, 1991] and thus focus small-scale thermally driven diapiric instabilities of plume material at the base of the lithosphere. This focused ascent of hot mantle material leads to rift initiation and breakup of the lithosphere at the craton boundary. In addition, the focusing mechanism observed in analogue models suggests that the along-strike segmentation of volcanic margins cannot be compared with the pattern of slow-spreading ridges. In the latter case, the segmentation is related to the focusing of accretion above small-scale instabilities of deep asthenospheric mantle [Lin *et al.*, 1990].

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