The Role of Magmatic Loads and Rift Jumps in Generating Seaward Dipping Reflectors on Volcanic Rifted Margins

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Abstract

The largest volcanic constructs on Earth are the seismically imaged seaward dipping reflector (SDR) units found offshore of many rifted continental margins, including most that border the Atlantic Ocean. Whether their formation requires large magnitude (i.e. 10s of km) of normal fault slip or results from the deflection of the lithosphere by the weight of volcanic flows is controversial. Though there is evidence for faulting associated with some SDRs, this paper considers the range of structures that can be produced by magmatic and volcanic loading alone. To do this an idealized mechanical model for the construction of rift-related volcanic flow structures is developed. Dikes open as plates move away from the center of a model rift and volcanic flows fill the depression produced by the load caused by dike solidification. The thin elastic plate flexure approximation allows a closed form description of the shape of both the contacts between flows and between the flows and underlying dikes. The model depends on two independent parameters: the flexure parameter, α , and the maximum isostatically supported extrusive layer thickness, w_0 . For reasonable values of these parameters the model reproduces the observed down-dip thickening of flows and the range of reflector dip angles. A numerical scheme using the analytic results allows simulation of the effect of temporal changes in the locus of magmatic spreading as well as changes in the amount

of volcanic infill. Either jumps in the location of the center of diking or periods with no volcanism result in separate units or “packages” of model SDRs, in which the flow-dike contact dips landward, consistent with observations previously attributed only to listric normal fault offset. When jumps in the spreading center are small (i.e. less than α) they result in thicker, narrower volcanic units on one side of a rift compared to those on the other side. This is similar to the asymmetric distributions of volcanic packages seen across some conjugate margins.

Keywords: Volcanic Margins; Continental Rifting; Seaward Dipping Reflectors; Flexure

1. Introduction

Most continental breakup events are preceded by massive extrusion of basalts comprising Large Igneous Provinces (LIPs) often located near the central part of a rift system (Courtillet et al., 1999). LIP volcanics were first identified on land where they typically are distributed on lateral dimensions of ~1000 km and have volumes of several million cubic kilometers (Coffin and Eldholm, 1994). The discovery of Seaward Dipping Reflectors (SDRs) greatly expanded the estimated lateral dimensions and volumes of rift-related volcanics. SDRs are sequences of seismically bright reflectors up to ~10 km thick and ~100 km wide were first identified on extension-parallel multi-channel seismic profiles crossing the rifted continental margins of the South and North Atlantic (Hinz, 1981, Mutter et al., 1982). The reflectors dip toward the ocean and bound layers that thicken down-dip. Although SDRs are usually buried under thick layers of sediment (**Figure 1**) drilling shows that they are made up of volcanic flows interleaved with sediments (Larson et al., 1994). Volcanic rifted margins (Figure 1A), defined by SDRs, are also underlain by regions with high seismic velocity, thought to indicate magmatic lower crust intruded at the time of SDR extrusion (Holbrook and Kelemen, 1993). SDRs have been found along one to two thousand kilometer-long sections of margins adjacent to LIPs and the total amount of intruded and extruded magma associated with a given breakup event can approach 10 million of cubic kilometers (Franke, 2013).

The finding of widespread SDRs has energized the debate about whether massive input of magma, likely resulting from mantle plumes, is needed to allow rifting. Before the recognition that most rifted margins are volcanic it was generally accepted that rifting occurred by amagmatic lithospheric stretching involving faulting of the brittle crust and

ductile flow of the warmer parts of the crust and mantle (e.g. McKenzie, 1978). Stretching explained how rifted continental margins subsided to produce space for thick piles of shallow water sediment. Yet the force needed for such tectonic stretching of lithosphere is estimated to be as much as ten times greater than the forces that may be available to drive rifting (Kusznir and Park, 1984; Naliboff et al., 2012). Since magma can allow lithospheric splitting, heating and thinning at far lower stress levels than needed for faulting (e.g. Rivalta et al., 2014), the inclusion of magma into rifting models offered a way to rift continents at modest force levels (Buck, 2004; Bialas et al., 2010).

A consequence of these models is that the tectonically accommodated fraction of plate separation should be small in a magma-rich environment, thereby precluding the formation of large-offset normal faults. Indeed, only very small offset (~10 m) normal faults are found in the magmatically robust regions of Iceland, primarily in areas of abandoned rift segments (e.g. Garcia et al., 2003, 2008; Siler and Karson, 2012). Many groups, however, have inferred a central role for faulting in the formation of SDRs based on the landward dip of the base of some seismically imaged volcanic units (Gibson and Love, 1989; Planke et al., 2000; Geoffroy, 2005; Quirk et al., 2014; and see Figure 1C). According to this ‘volcanic faulting model’ extrusive basalts fill the depression created by 10’s of kilometers of offset of a normal fault bounding a half-graben, in much the same way syn-rift sediment fills magma-poor extensional basins (Figure 2A). Geoffroy (2005) has marshaled many observations that favor a role for normal faults that dip landward in SDR formation, including the observation of landward dipping faults cutting in sections of basalt flows exposed on land in Eastern Greenland. However, a landward dip is opposite to what is seen for most rift-bounding and continental margin normal

78 faults (Sibuet, 1992) and expected from analog (Brun, 1999) and numerical rift models
79 (Huisman et al., 2005). A numerical model by Geoffroy et al. (2015) produces faults
80 that dip landward but only by assuming a complex initial lithospheric structure and
81 evolution.

82 Further evidence of a role for tectonic faults in volcanic margin formation is inferred
83 from the asymmetric distribution of some SDRs packages across conjugate rifted
84 margins. For example, along a portion of the South Atlantic margins Becker et al. (2016)
85 identify two to three SDR packages on seismic lines crossing each conjugate margin (see
86 Figure 1D and 1E). Up to a hundred kilometers of slip on lithosphere-scale normal
87 “detachments” has been suggested as a way to produce thicker and narrower volcanic
88 units on the Argentinean side compared to the South African side (Becker et al., 2016).
89 Similar large-offset detachments have been suggested to explain the structure of other
90 volcanic margins (e.g. Pindell et al., 2014).

91 Though faulting may play a role in the development of some SDRs, the object of this
92 paper is to consider what volcanic structures can be produced during rifting without
93 large-offset faults. The quantitative mechanical model developed here builds on
94 previous, largely kinematic, models of the development of SDRs. An analytic
95 description of the response of the lithosphere to the loads of dike solidification and
96 volcanic flows is developed. This description is then used to show that jumps of the
97 center of spreading and volcanism can result in volcanic flow unit geometries that are
98 similar to those observed at some volcanic rifted margins. The formulation developed
99 here can form the basis of more rigorous numerical models of the development of
100 volcanic rifted margins.

2. Previous Work on the Formation of SDRs due to Volcanic Loading

The earliest interpretation of SDRs was based on observations of the structure of massive basalt flows that dip toward the subaerial spreading center on Iceland (Mutter et al., 1982). Glacially carved fjords provide a window into these volcanic piles and show that the layers dip toward the center of the island. This led (Bodvardsson and Walker, 1964) to develop a ‘volcanic loading model’ that required plate spreading even before plate tectonics had been widely accepted. Their model assumes local isostatic compensation of volcanic flows fed by dikes opening repeatedly close to a plate-spreading center. The flows are assumed to be thickest close to the axis of diking and to have a triangular cross-section. This explains the increasing dip of volcanic flows with depth as well as the down-dip thickening of volcanic flows seen on Iceland and for SDRs.

Based on the qualitative arguments of Bodvarsson and Walker (1964), Palmasson, (1980) developed a kinematic model for the development of multiple volcanic flows assuming local isostasy and a particular geometry of volcanic extrusion. To go from kinematic to dynamic models requires specification of loads that produce surface lowering and so space for extrusives to fill. Four ideas about the origin of such loads have been suggested. The effect of the load of a migrating volcanic construct on lithospheric flexure was treated by Watts and Cox (1989) to explain the geometry of down-dip thickening of basalt flows seen for the Deccan Traps. Bastow and Keir (2011) suggest that the thinning of the ductile lower crust by stretching causes isostatic subsidence that allows infilling by extrusives. Pindell et al (2014) suggest that emptying of a magma chamber should pull down the surface. Finally, Corti et al. (2014) show that intrusion of dense mafic rocks into a felsic crust can flex the upper crustal lithosphere and make space for

the kinds of volcanic structures observed in parts of the East African Rift System. All of these processes may act at a volcanic margin, but are difficult to incorporate in a simple, general model with a minimum of free parameters. As in Corti et al. (2014) the model described below focuses on the effect of mafic dike intrusions that are seen to accommodate plate separation at volcanic rifts. However, in the new model flexure is driven by the density increase on solidification of series of basaltic dikes that fill in the space produced by plate separation. Combined with simplifying assumptions about magma extrusion and thin plate flexure, this allows derivation of an analytic description of the geometry of resulting volcanic structures.

3. Analytic model of ‘Volcanic Flexure’

When a basaltic dike solidifies the related density increase should load the lithosphere and pull down the surface, creating space that can be filled in by extrusives. Derivation of an analytic description of the geometry of extrusive flows resulting from such loading and infill follows an approach used to treat dike solidification at a mid-ocean ridge, without volcanic flows, that results in an axial high similar to those seen at fast spreading ridges (Buck, 2001, Shah and Buck, 2001, 2004).

Fluid magma of density ρ_f in a dike at the axis of rifting or spreading, at $x = 0$, is taken to rise to its level of local isostatic compensation (Figure 3a). As the dike solidifies it becomes part of the adjacent lithospheric plates and as the density of the solid dike increases to ρ_d it loads and deflects the plates (Figure 3b). If the dike height h_d is much smaller than its along-axis length then a two-dimensional cross-sectional model, involving vertical depth z and horizontal distance x from the center of dike injection, is

146 justified. A dike of half-width δx that solidifies from fluid magma will produce a load
 147 per unit rift length on each plate of:

$$\delta V_0 = \delta x g h_d \Delta \rho_d \quad (1)$$

148 where $\Delta \rho_d = (\rho_d - \rho_f)$ and g is the acceleration of gravity. Extrusives are assumed to fill
 149 in and amplify the depression produced by the load of the solidified dike (Figure 3c).
 150 Two approximations are needed to allow derivation of an analytic expression for the
 151 deflection of the lithosphere by combined dike and extrusive loads. First, the lithosphere
 152 is treated as a thin elastic plate (Watts, 2001) and, second, the region above the flexing
 153 plate is treated as if it were an inviscid fluid with the density of the infilling material.
 154 Assuming the plate has a laterally uniform rigidity D the relevant form of the thin-plate
 155 flexure equation for vertical deflection w is then:

$$D \frac{d^4 w}{dx^4} + g \Delta \rho_s w(x) = 0 \quad (2)$$

156 where the flexure parameter $\alpha = (4D/g\Delta\rho_s)^{1/4}$ and g is the acceleration of gravity. $\Delta\rho_s$,
 157 the compensation density contrast, is the difference between the density of substrate
 158 material displaced by the deflected plate ρ_s and the density of the infilling material ρ_i , so
 159 that $\Delta\rho_s = (\rho_s - \rho_i)$.

160 The lithosphere at the axis, $x = 0$, is broken by the dike so that the applied bending
 161 moment and the effective plate rigidity there are zero. Then the dike solidification load
 162 given by equation 1 applied at the axis produces vertical surface deflections:

$$\delta w(x) = \delta w_0 \exp\left(-\frac{x}{\alpha}\right) \cos\left(\frac{x}{\alpha}\right) \quad (3)$$

163 where $\delta w_0 = 2\delta V_0/(\alpha g \Delta \rho_s) = 2\delta x h_d \Delta \rho_d/(\alpha \Delta \rho_s)$.

164 Following Buck (2001) it can be shown that the repeated intrusion of an infinite number
165 of dikes with infinitesimal width results in relief of the tops of the dikes given by
166 integration of equation 3 from 0 to x , yielding:

$$w_t(x) = w_0 \left\{ \exp\left(-\frac{x}{\alpha}\right) \left[\sin\left(\frac{x}{\alpha}\right) - \cos\left(\frac{x}{\alpha}\right) \right] + 1 \right\} \quad (4)$$

167 where $w_0 = h_d (\Delta \rho_d / \Delta \rho_s)$. The position of the top surfaces of individual extrusive flows
168 predicted by this model can also be described analytically assuming that the flow surface
169 originates at zero height (i.e. at the height of the top of the fluid dike at the axis). Then
170 the depth w_s of a flow surface is given integration of equation 3 from $x-x_0$ to x , giving:

$$w_s(x) = w_0 \left\{ \exp\left(-\frac{x}{\alpha}\right) \left[\sin\left(\frac{x}{\alpha}\right) - \cos\left(\frac{x}{\alpha}\right) \right] \right. \\ \left. - \exp\left(-\frac{(x-x_0)}{\alpha}\right) \left[\sin\left(\frac{(x-x_0)}{\alpha}\right) - \cos\left(\frac{(x-x_0)}{\alpha}\right) \right] \right\} \quad (5)$$

171 where x_0 is the horizontal distance from the axis to the contact between the flow and the
172 underlying dike. Figure 4 shows that this model produces a geometry of extrusive layers
173 that is similar to that observed for a sequence of flows in an SDR. The non-dimensional
174 axes of that plot emphasize that the results depend on two length scales: the flexure
175 parameter α and w_0 .

176 The horizontal wavelength of lithospheric bending is set by the flexure parameter α . For
177 the thin elastic plate approximation used in the present model the rigidity $D = Eh^3/12(1-\nu^2)$
178 where E is Young's Modulus, h is the effective elastic thickness of the layer and ν is

Poisson's ratio. For $h = 5$ km, $E = 5 \times 10^{10}$ Pa, $\nu = 0.25$ and $\Delta\rho_s = 200$ kg/m³ the flexure parameter would be ~ 33 km.

The vertical scale of the volcanic pile is controlled by w_0 , which is equivalent to the thickness of extrusives in local isostatic equilibrium with the fluid dike at the axis. In an off-axis column extrusive infill of density ρ_i overlies a solidified dike layer of height h_d and density ρ_d while an on-axis column consists of a fluid dike of height h_d and density ρ_f . The on- and off-axis regions are taken to float on a compensating layer of density ρ_c that is sufficiently weak that lateral pressure gradients are negligible. Having the same overburden pressure at the base of the on- and off-axis columns implies that $w_0 = h_d(\rho_d - \rho_f)/(\rho_c - \rho_i) = h_d(\Delta\rho_d/\Delta\rho_c)$, consistent with the flexural model derivation. This illustrates that it is the increase in density of the solid dikes relative to the fluid dike at the axis that results in a finite thickness of the extrusive layer.

The vertical deflections depend strongly on the density of the infilling material. Assuming $h_d = 10$ km, $\Delta\rho_d = 200$ kg/m³, with zero density air as the infilling material (i.e. for $\rho_i = 0$) and taking $\rho_c = 3000$ kg/m³ then w_0 would be 667 m. With the same parameters except with lava density as the infilling material (i.e. with $\rho_i = 2800$ kg/m³) then w_0 would be 10 km.

4. The effect of rift axis jumps

The analytic volcanic flexure model produces the same distribution of SDR flow units on both sides of a model rift (Figures 3 & 4). The production of asymmetry may be accomplished by offsets in the position of the center of dikeing such as those seen at

200 spreading centers. For example, several offsets or jumps in the location of Icelandic
201 spreading centers are well documented and one is ongoing in southern Iceland where the
202 eastern rift zone is replacing the dying western rift zone (Hardarson et al., 1997).
203 Submarine ridge propagation produces offsets in the position of the spreading center on
204 length scales that are usually smaller than the jumps seen on Iceland and they can result
205 in asymmetries and gaps in the magnetic record of seafloor age (Hey, 1979). Several
206 processes may cause spreading center jumps including plume ridge offset (e.g
207 Mittelstaedt et al., 2011) as well as shifts in the spreading direction (e.g. Hey, 1979). An
208 illustration of how jumps can produce separate SDR packages in the context of volcanic
209 flexure is shown in Figure 2B.

210 Incorporation of such spreading center jumps requires numerical summation of the effect
211 of multiple small dike-extrusion events. To do this a uniform grid of points a distance δx
212 apart in the horizontal x-direction are specified where cumulative vertical deflections
213 calculated using equation 2 are tallied. If each dike has a width of $2\delta x$ then dike opening
214 offsets all grid points by one unit. A model time-step involves adding the incremental
215 deflection induced by dike solidification and volcanic infilling, and then laterally
216 offsetting the grid point values by δx .

217 Two length scales have to be specified to simulate jumps: the width, Δx_{ext} , of the
218 extrusive packages on each side of the rift between jumps and the horizontal distance of a
219 jump, Δx_{jump} . Jumps are accomplished by reassignment of the location of the numerical
220 center of spreading relative to the marker points defining extrusive layers. To illustrate
221 the structure of the extrusives a new set of markers is added periodically at a depth = 0,

approximating the surface of an extrusive flow. Jumps can result in separate packages of flows that may overlap. Results of a model case with particular values of Δx_{ext} and Δx_{jump} are shown in Figure 5B.

5. The effect of gaps in extrusion

Separate SDR units could also be produced if there were periods with dike accommodated spreading but little or no extrusion. The existence of gaps in extrusion at spreading centers is more speculative than the existence of jumps, but gaps in extrusion are included here for completeness. There is evidence from the spreading segment at 16°N on the East Pacific Rise (EPR) that the volcanic flux may first increase and later decrease after a ridge jump. There the axial high formed after a ridge jump occurred (at 150 to 300 Ka) is much higher than is typical for fast spreading ridges and apparently higher than the high that was abandoned (Carbotte et al., 2000; Shah and Buck, 2005).

To simulate gaps an extrusive phase of width Δx_{ext} is followed by a phase of width Δx_{gap} with no extrusive loading during which an infill density of zero is specified. Figure 5C shows a case with 3 overlapping packages of extrusive layers produced by assuming alternating extrusive and gaps phases and markers are not added during a gap phase. If the gap phase is wider than about twice the flexure parameter the packages do not overlap.

6. Discussion

A cursory comparison of model predictions (Figures 4 and 5) with the seismic reflection images of SDRs in Figure 1 shows that the model produces the general shape and

inferred internal structure of for a range of SDR packages. The analytic version of this model allows clear illustration of the relations between model parameters and predictions that can be tested with observations. For example, the maximum slope of the base of the flows is $2w_o/\alpha$, and the dips equal $\arctan(2w_o/\alpha)$. For $w_o = 5\text{-}10$ km and a flexure parameter α of 20-60 km the predicted maximum dips range from 10 to 45 degrees which is comparable to or wider than the observed range.

The predicted thickness of the extrusive pile depends on the height of the intruded dike and the density at a depth of compensation of the loads. The height of dike opening should be close to the thermally-controlled thickness of brittle lithosphere in a rift and this is difficult to estimate for a given rift. Based on the observed depths of seismicity during dike intrusion events along the volcanic rich, slowly spreading ridge in Iceland (Einarsson and Brandsdottir, 1980) a reasonable value of the dike height h_d is 10 km. If the lower crust is very weak then compensation can occur there, otherwise compensation should occur in the mantle. For $h_d = 10$ km and compensation of the loads in the lower crust, with a substrate density ρ_s of 3000 kg/m³, the model gives $w_o \sim 10$ km, assuming an infill density ρ_i of 2800 kg/m³ and an increase in dike density on solidification $\Delta\rho_d$ is 200 kg/m³. With the same parameters except with compensation in mantle, with $\rho_s = 3300$ kg/m³, the SDR package would be as thin as 4 km. Thus, thick SDRs may imply that the lower crust was weak enough to flow easily at the time of their emplacement.

To produce the asymmetry in the thickness and width of SDRs requires several jumps in location of the center of dikeing and that the distance of the jumps is smaller than the

flexure parameter α . Figure 6 compares a particular model with three small jumps to interpreted sections of SDRs for conjugate South Atlantic margins.

Jumps that are larger than α should result in thicker extrusive piles on both sides of the new center of dikeing. Such large-scale jumps have happened on Iceland (Hardarson et al., 1997) and this model makes the potentially testable prediction of thicker extrusives on both sides of the region where a new center of spreading formed, as shown in Figure 7.

The large (~ 100 km) jumps on Iceland are likely related to migration of the spreading center away from a plume, as noted earlier. The small jumps required to produce asymmetries across volcanic rifted margins in the context of this model may be related to offsets of the pole of opening. The geometry of some rifts and margins seems to indicate such offsets. For example, the present direction of opening of the Red Sea is not orthogonal to the overall trend of the rift change (Mohriak and Leroy, 2012). If the early rift opening weakened the lithosphere and set the rift trend then the present geometry might reflect a shift in the opening direction.

Either jumps in the spreading center or gaps in extrusion can lead to separate SDR packages that look similar to those predicted by faulting models. The fact that a combination of dike solidification and volcanic loads can produce many of the features that have been attributed to faulting does not necessarily mean that faulting plays no role in the formation of SDRs. There is observational evidence, both on land and from seismic records (Geoffroy, 2005) that faulting does accompany volcanic flow structures. One of the most promising ideas about how this could happen is the suggestion by Quirk

et al. (2014) that an axial horst, underlain by low-density partial melt may be bordered by outward dipping normal faults. Two-dimensional numerical simulations of this conceptual are being formulated.

More work needs to be done to justify the assumptions and simplifications of the present model. To allow an analytic description, extrusives are assumed to be present not only in the flexural depression near the spreading axis but also above the flanking flexural bulge. This approximation leads to a larger flexural bulge than would exist if extrusives only filled in the flexural depression. Numerical treatment of a thin plate elastic model with extrusive infill only to the level of the spreading axis show that simplifying approximation changes the predicted dip and thickness of the beds by only a few percent. Also, there has to be some topographic slope for lava to flow away from the axis, but a zero slope is assumed in the model. One reason is that flows in continental flow basalt provinces form with very low surface slopes, as low as 0.01 (Keszthelyi and Self, 1998). Assuming a finite slope greatly complicates the model and adds a poorly constrained parameter.

Another limitation of the thin plate flexure model is that it can only describe vertical deflections and so the rotations of vertical planes that should occur as a plate is bent are not treated. Rotations of the initially horizontal flows are well approximated and dikes should be orthogonal to the flows at their contact as indicated in the model schematic of Figure 3. The precise rotation of dikes and flows can only be accurately treated with a two-dimensional model.

307 In addition, the finite strength of the lithosphere can result in yielding of the top and
308 bottom of a bending plate and this reduces the plate rigidity and so the wavelength of
309 flexure decreases as loads and bending increase (e.g. Bodine et al., 1981). The analytic
310 model predicts curvatures of order 10^{-5} m^{-1} and these would suggest elastic stresses of
311 greater than a GPa and these would certainly cause brittle failure of parts of the
312 lithosphere. Fully, two-dimensional elasto-plastic models of lithospheric bending under
313 large loads suggest that the effective elastic thickness of a finite-strength brittle layer may
314 only be a fraction of its true thickness (e.g. Lavier et al., 2000). Thus, the appropriate
315 values of the flexure parameter should be much smaller than for a purely elastic plate of a
316 given thickness.

317 A two-dimensional version of this model with more realistic lithospheric rheologies
318 could better treat the strong bending that should result in brittle yielding of the
319 lithosphere via distributed faulting. One question worth study is whether bending related
320 faulting could explain the observed small offset faulting seen in the coastal flexure zone
321 of Greenland and previously used as evidence for a key role of faulting in SDR formation
322 (e.g. Geoffroy, 2005). Another key question to be addressed with a more complete model
323 is the rapid subsidence of a volcanic passive margin (Pindell et al., 2014). One
324 possibility to be tested is whether a reduction in the rate of magma input would lead to
325 the kind of crustal thinning during continued rifting that has been suggested by Bastow
326 and Keir (2011). However, the main conclusion of the present work, that some volcanic
327 margin structures can be produced without faulting, is unlikely to be altered by more
328 advanced work.

Some questions about SDR formation will require consideration of very large-scale geodynamic processes. For example, at many volcanic rifted margins there is evidence of separate inner and outer SDRs. One can speculate that the early (inner) phase of SDR formation may reflect melt production by active flow in a plume or plume head that is focused in a region of LIP formation. Transport of this early forming magma from the locus of LIP formation to erupt along the rift axis might be accomplished via dikes. The later (outer) SDRs might reflect passive upwelling of hotter-than-normal mantle and vertical transport of resulting melts in the way envisioned by White and McKenzie (1989).

7. Conclusions

Numerous authors suggest that many kilometers of normal fault offset are needed to provide the space for the thick volcanic flows comprising the seaward dipping reflector sequences that define volcanic rifted margins (Gibson and Love, 1989; Planke et al., 2000; Geoffroy, 2005; Quirk et al., 2014). This view challenges our understanding of the basic mechanics of magma intrusion during rifting. Theory predicts that if sufficient low-viscosity basaltic magma is always available for extrusion that extensional stresses should not reach the level needed for lithosphere-cutting normal faults (e.g. Rivalta et al, 2015). Observations in areas of massive sub-aerial basaltic volcanism, like Iceland, suggest that ridge-ward dipping volcanic flow structures, similar to SDRs, can form without major normal fault offsets (e.g. Bodvarsson and Walker, 1964).

This paper shows that several SDR-related structures, often attributed to faulting, can be produced by magmatic and volcanic loading alone. In the quantitative “volcanic flexure” model derived here the space for volcanic flows results from the load of magmatic dikes

solidifying and densifying after emplacement at a rift center (Figure 2). This description of the mechanics of rift magmatism builds directly on the kinematic models of volcanic flow loading (Bodversson and Walker, 1964, Palmasson (1980) that were inspired by observations of ridge-ward thickening sequences of lava flows in Iceland.

Approximations, including treating lithospheric deflection in terms of thin elastic plate flexure, allow an analytic description of the shape of the contacts between flows and between the flows and underlying dikes. The model geometry depends on two independent length scales: the flexure parameter, α , and the maximum isostatically supported extrusive layer thickness, w_0 (Figure 3). For reasonable values of these parameters the model reproduces the observed range of flow dip angles in SDRs.

A numerical scheme using these analytic results allows simulation of the effect of temporal changes in the locus of magmatic spreading and the level of volcanic infill. Either jumps in the location of the diking center or periods with no volcanism result in separate units or “packages” of model SDRs, in which the flow-dike contact dips landward, consistent with observations previously attributed only to listric normal fault offset. When jumps are large compared to the flexural parameter the model reproduces the pattern of flow unit dips seen in abandoned spreading centers in Iceland (Hardarson et al.; 1998; Figure 7). Several small magnitude jumps in the same direction produce the kind of asymmetric SDRs seen across conjugate rifted margins of the South Atlantic (Figure 6) and previously attributed to lithospheric scale detachment faulting (Becker et al., 2016).

Magma intrusion and volcanism affect many rifts and continental margins but are ignored in most numerical models of lithospheric extension, including recent studies (e.g.

Svartman Dias et al., 2015; Brune et al., 2016). Only a few recent models include idealized simulations of magma intrusion in long-term lithospheric extension models (e.g. Bialas et al., 2010; Keller et al. 2013). The present work may stimulate development of more complete numerical treatments of effects of rift magmatism, since such models can be “benchmarked” against the analytic results given here.

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References

- Barton, A.J., and R. S. White (1997) Volcanism on the Rockall continental margin: Geol. Soc. London, 154, 531–536, doi:10.1144/gsjgs.154.3.0531.
- Bastow, I., and D. Keir (2011) The protracted development of the continent-ocean transition in Afar: Nature Geoscience, v. 4, p. 248–250, doi:10.1038/ngeo1095.
- Becker, K., D. C. Tanner, D. Franke, and C. M. Krawczyk (2016), Fault-controlled lithospheric detachment of the volcanic southern South Atlantic rift, Geochem. Geophys. Geosyst., 17, doi:[10.1002/2015GC006081](https://doi.org/10.1002/2015GC006081).
- Bialas, R, W. R. Buck and R. Qin (2010) How Much Magma is required to rift a continent? Earth and Planet Sci. Lett., 292, 68-78, 2010.
- Bodine, J.H., M.S. Steckler and A.B. Watts (1981) Observations of Flexure and the Rheology of the Oceanic Lithosphere, J. Geophys. Res., 86, 3695-3707.

396 Bodvarsson, G. & Walker, G. P. L. (1964) Crustal drift in Iceland, *Geophys. J. R. astr.*
397 *Soc.*, 8, 285-300.

398 Brun, J. P. (1999), Narrow rifts versus wide rifts: Inferences for the mechanics of rifting
399 from laboratory experiments, *Phil. Trans. R. Soc. Lond. A*, 357, 695-712.

400 Brune, S., S.E. Williams, N.P Butterworth and D. R. Müller (2016) Abrupt plate
401 accelerations shape rifted continental margins, *Nature*, doi:10.1038/nature18319.

402 Buck, W. R.(2001), Accretional curvature of lithosphere at magmatic spreading produces
403 an elastically supported axial high, *J. Geophys. Res.*, 106, 3953-3960.

404 Buck, W. R. (2004) Consequences of asthenospheric variability on continental rifting, in
405 *Rheology and deformation of the lithosphere at continental margins*, editors G.D.
406 Karner, B. Taylor, N.W. Driscoll and D.L. Kohlstedt, Columbia University Press, 1-
407 31.

408 Carbotte, S., A. Solomon, and G. Ponce-Correa (2000), Evaluation of morphological
409 indicators of magma supply and segmentation from a seismic reflection study of the
410 East Pacific Rise 15°30'–17°N, *J. Geophys. Res.*, 105, 2737– 2759.

411 Coffin, M. F., and O. Eldholm (1994), Large igneous provinces: Crustal structure,
412 dimensions, and external consequences, *Rev. Geophys.*, 32(1), 1–36,
413 doi:[10.1029/93RG02508](https://doi.org/10.1029/93RG02508).

414 Corti, G., A. Agostini, D. Keir, J. Van Wijk, I. D. Bastow, and G. Ranalli (2015) Magma-
415 induced axial subsidence during final stage rifting: Implications for the development
416 of seaward-dipping reflectors, *Geosphere*, 11, DOI: 10.1130/GES01076.1.

417 Courtillot, V., C. Jaupart, I. Manighetti, P. Tapponnier, J. Besse (1999) On causal links

418 between flood basalts and continental break-up, *Earth Planet. Sci. Lett.*, 166, 177–
 419 195.

420 Einarsson, P. and B. Brandsdottir (1980) Seismological evidence for lateral magma
 421 intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland, *J.*
 422 *Geophys.*, 47, 160– 165.

423 Franke, D. (2013), Rifting, lithosphere breakup and volcanism: Comparison of magma-
 424 poor and volcanic rifted margins, *Marine and Petroleum Geology* 43, 63-87,
 425 doi:10.1016/j.marpetgeo.2012.11.003.

426 Garcia, S., N. O. Arnaud, J. Angelier, F. Bergerat, and C. Homberg (2003), Rift jump
 427 process in northern Iceland since 10 Ma from $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology, *Earth*
 428 *Planet. Sci. Lett.*, 214, 529 – 544.

429 Garcia, S., J. Angelier, F. Bergerat, C. Homberg, and O. Dauteuil (2008) Influence of rift
 430 jump and excess loading on the structural evolution of Northern Iceland, *Tectonics*,
 431 27, TC1006, doi:10.1029/2006TC002029, 2008.

432 Geoffroy, L., 2005, Volcanic passive margins: *Comptes Rendus Geoscience*, 337, 1395–
 433 1408, doi: 10.1016 /j .crte.2005.10.006.

434 Gibson, I.L., D. Love (1989) A listric fault model for the formation of the dipping
 435 reflectors penetrated during the drilling of hole 642E, ODP Leg 104, in: O. Eldholm,
 436 J. Thiede, E. Taylor, et al. (Eds.), *Proc. ODP, Sci. Results*, 104, 979–983.

437 Hardarson, B.S., Fitton, J.G., Ellam, R.M., Pringle, M.S. (1997) Rift relocation - a
 438 geochemical and geochronological investigation of a paleo-rift in northwest Iceland.
 439 *Earth Planet. Sci. Lett.* 153, 181–196.

440 Hey, R. N. (1979), Evidence for spreading center jumps from fine-scale bathymetry and
 441 magnetic anomalies near the Galapagos Islands, *Geology*, 7, 504– 506.

442 Hinz, K. (1981) A hypothesis on terrestrial catastrophes: wedges of very thick oceanward
 443 dipping layers beneath passive margins – Their origin and palaeoenvironment
 444 significance, *Geol. Jahrb.*, 22, 345–363.

445 Holbrook, W. S., and Kelemen, P. B. (1993) Large igneous province on the U.S. Atlantic
 446 margin and implications for magmatism during continental breakup: *Nature*, 364,
 447 433–436.

448 Hopper, J. R., J. T. Dahl, et al. (2003) Structure of the SE Greenland margin from seismic
 449 reflection and refraction data; implications for nascent spreading center subsidence
 450 and asymmetric crustal accretion during North Atlantic opening." *J. Geophys. Res.*,
 451 108: 2269.

452 Huismans, R. S., S. J. H. Buiters, and C. Beaumont (2005), Effect of plastic-viscous
 453 layering and strain softening on mode selection during lithospheric extension, *J.*
 454 *Geophys. Res.*, 110, B02406, doi:10.1029/ 2004JB003114.

455 Keller, T., D. A. May, and B. J. P. Kaus (2013) Numerical modelling of magma
 456 dynamics coupled to tectonic deformation of lithosphere and crust, *Geophys. J. Int.*,
 457 195, 1406–1442.

458 Keszthelyi, L. and S. Self (1998) Some physical requirements for the emplacement of
 459 long basaltic lava flows *J. Geophys. Res.* 103,2156-2202 doi.org/10.1029/98JB00606.

460 Kusznir N. J. and R. G. Park (1984) Intraplate lithosphere deformation and the strength of
 461 the lithosphere, *Geophys. J. Roy. Astron. Soc.*, 79, 513–538.

462 Larsen, H.C., et al. (1994) Proceedings of the Ocean Drilling Program, Initial Reports, 152, Ocean
 463 Drill. Program, College Station, Tx.

464 Lavier, L., W. R. Buck and A.N. B. Poliakov (2000), Factors controlling normal faulting
 465 in ideal elastic-plastic layers, *J. Geophys. Res.*, 105, 23,431-23,442.

466 McKenzie, D. P. (1978) Some remarks on the development of sedimentary basins, *Earth*
 467 *Planet. Sci. Letts.*, 40, 25-32.

468 Mittelstaedt, E., G. Ito, and J. van Hunen (2011) Repeated ridge jumps associated with
 469 plume-ridge interaction, melt transport, and ridge migration, *J. Geophys. Res.*, 116,
 470 doi:10.1029/2010JB007504.

471 Mohriak, W. U. and S. LeRoy (2012) Architecture of rifted continental margins and
 472 break-up evolution: insights from the South Atlantic, North Atlantic and Red Sea—
 473 Gulf of Aden conjugate margins, *Geol. Soc. Lond., Sp. Pubs.*, 369:497-535, doi:
 474 10.1144/SP369.17

475 Mutter, J. M. Talwani, P.L. Stoffa (1982) Origin of seaward-dipping reflectors in oceanic
 476 crust off the Norwegian margin by ‘subaerial sea-floor spreading’, *Geology*, 10, 353-
 477 357.

478 Naliboff, J.B., Lithgow-Bertelloni, C., Ruff, L.J. & de Koker, N. (2012) The effects of
 479 lithospheric thickness and density structure on Earth’s stress field, *Geophys. J. Int.*,
 480 188, 1–17.

481 Neilsen, T. F. D. and C. K. Brooks (1981) The E Greenland rifted continental margin; an
 482 examination of the coastal flexure, *Journal of the Geol. Soc. London*, 138, 559-568.

483 Palmason, G. (1980) A continuum model of crustal generation in Iceland: kinematic

aspects, *J. Geophys.*, 47, 7–18.

Pindell, J., Graham, R., and Horn, B. (2014) Rapid outer marginal collapse at the rift to drift transition of passive margin evolution, with a Gulf of Mexico case study: *Basin Research*, 26, 1–25, doi:10.1111/bre.12059.

Planke, S., P.A. Symonds, E. Alvestad, J. Skogseid (2000) Seismic volcanostratigraphy of large-volume basaltic extrusive complexes on rifted margins, *J. Geophys. Res.*, 105, 19335–19352.

Quirk, D.G., A. Shakerley, and M.J. Howe (2014), A mechanism for construction of volcanic rifted margins during continental breakup, *Geology*, 42, 1079–1082, doi:10.1130/G35974.

Rivalta E., B. Taisne , A.P. Bungler, R.F. Katz (2014) A review of mechanical models of dike propagation: schools of thought, results and future directions, *Tectonophysics* 638 (2015) 1–42, <http://dx.doi.org/10.1016/j.tecto.2014.10.003>.

Shah, A. and W. R. Buck (2003) Plate bending stresses at axial highs, and implications for faulting behavior, *Earth Planet. Sci. Letts.*, 211, 343–356.

Shah, A. and W. R. Buck (2006) The rise and fall of axial highs during ridge jumps, *J. Geophys. Res.*, 111, B08101, doi:10.1029/2005JB003657.

Sibuet, J. C. (1992) New constraints on the formation of the non-volcanic continental Galicia–Flemish Cap conjugate margins, *J. Geol. Soc. Lond.*, 149, 829–840.

Siler, D. L., and J. A. Karson (2012), Subvolcanic subsidence and caldera formation during subaerial seafloor spreading in Iceland, *Geol. Soc. Am. Bull.*, 124(7/8), 1310–1323, doi:10.1130/B30562.1.

506 Svartman Dias, A. E., L. L. Lavier and N. W. Hayman, 2015, Conjugate rifted margins
507 width and asymmetry: The interplay between lithospheric strength and
508 thermomechanical processes, *J. Geophys. Res.*, 120, 2015JB012074, doi:
509 10.1002/2015JB012074.

510 Watts, A. B. (2001) *Isostasy and Flexure of the Lithosphere*, Cambridge University
511 Press, 458 p.

512 Watts, A. B. and K. G. Cox (1989) The Deccan Traps: An interpretation in terms of
513 progressive lithospheric flexure in response to a migrating load, *Earth Planet. Sci.*
514 *Lett.*, 93, 85-97.

515 White, R. S., and McKenzie, D. (1989) Magmatism at rift zones: The generation of
516 volcanic continental margins and flood basalts: *Journal of Geophysical Research*, 94,
517 7685–7729, doi:10.1029/JB094iB06p07685.

Figure Captions

Figure 1. (A) Location map of well documented volcanic rifted margins (red lines from Geoffroy, 2005) and the locations of the multichannel seismic (MCS) reflection lines shown below the map and in Figure 6. B shows MCS data from the East Greenland Margin (from Hopper et al. (2003). C shows MCS data from the Exmouth Plateau of Australia showing the 3 separate overlapping SDRs ‘units’ indicated with colored shading (from Planke et al., 2000). The top of volcanic sequence is labeled ‘t’, prominent reflectors dividing the sequence into sub-units are labeled ‘r’. D show MCS lines on the Argentinean side of the South Atlantic and E shows similar data from the conjugate margin the South Africa with lines and letters representing interpreted boundaries between flow units (from Becker et al., 2016).

Figure 2. Illustration of two models for producing separate seaward dipping reflector “packages” on a volcanic rifted margin. (A) shows the fault infill model (after Geoffroy, 2005) and (B) the volcanic flexure model, both described in the text. Light lines denote volcanic flow tops and each color is meant to represent a sequence of flows formed during a given time interval. The nail on the right of the panels indicates a frame of reference in which the right margin is fixed.

Figure 3. Illustration of the volcanic flexure model for the development of SDRs. Panel (A) shows a dike filled with fluid magma open between two lithospheric plates. The density increase caused by solidification of the dike in (B) loads and bends the plates.

Panel (C) shows how extrusion of magma amplifies the deflections. The results of 2 dike intrusion plus extrusion events is shown schematically in (D). The size of dikes and flows is exaggerated in A through D. (E) shows the results of many small dike intrusion plus extrusion events and emphasizes that the model depends on two length scales, α and w_0 , described in the text.

Figure 4. Plots of the analytically computed positions of a series of volcanic layer tops (grey lines) as well as the location of the dike-flow contact (a black line) for the analytic volcanic flexure model described in the text.

Figure 5. Results of three model cases that illustrate some of the diversity of SDRs that can result from volcanic flexure. As in Figures 1 and 2 the lines dipping toward the spreading axis, where distance = 0, represent the tops of flow units. The same values for flexural parameter ($\alpha = 20\text{km}$) and characteristic deflection ($w_0 = 10\text{km}$) were used for all cases. (A) shows the model flow geometry after a total of 60 km of plate separation (30 on each rift side) with a fixed location of the center of dike intrusion with extrusion constantly filling the flexural low. (B) shows asymmetry in the horizontal distribution of SDR packages due to a sequence of periods each with 40 km of spreading followed by a 10 km leftward jump in the spreading center. (C) shows how periods with 40 km of spreading and extrusion followed by a “gap” with 20 km of dike accommodated spreading without volcanic extrusion, can result in distinct SDR packages.

Figure 6. A and B show interpreted MCS lines (with depth sections based on refraction

data) on conjugate margins of the South Atlantic and letters represent boundaries between flow units (from Becker et al., 2016). The interpreted SDR packages are thicker and narrower on the western (South American) margin than the eastern (African) margin. C Shows the predicted extrusive layer geometry resulting from the volcanic flexure model described in the text. For that model case the flexure parameter α is 20 km, the characteristic volcanic layer thickness w_0 is 12 km. The parameters on the lower right side of the plot describing the sequence of 3 periods normal extrusion for 10 km of spreading followed by a gap of 5 km with no extrusion then a 5 km jump to the left.

Figure 7. Map at top shows the positions of present and past (dotted lines) regions of volcanism and focused plate separation for Iceland (from Hardasson et al 1997). The present Northern, Middle, Western and Eastern Rift Zones are labeled accordingly. The approximate age range of past spreading and volcanic activity are shown adjacent to the dotted lines. The sequence of model results shown below indicate the effect of several large jumps (jump distance $> \alpha$).

Illustration of the volcanic flexure model for the development of SDRs. Panel (A) shows a dike filled with fluid magma open between two lithospheric plates. The density increase caused by solidification of the dike in (B) loads the bends the plates. Panel (C) shows how extrusion of magma amplifies the deflections. The results of 2 dike intrusion plus extrusion events is shown schematically in (D). The size of dikes and flows is exaggerated in A through D. (E) shows the results of many small dike intrusion plus extrusion events and emphasizes that the model depends on two length scales, α and w_0 .

(A) Volcanic Rifted Margins

(B) East Greenland Margin

(C) Western Australian Margin

(D) South American Margin

(E) African Margin

Figure 1.

A. Faulting with Volcanic Infill

B. Volcanic Flexure with Jumps

Figure 2

Fluid magma filled dike intrudes crustal lithosphere

Solidification of dike loads lithosphere

Extrusives further load lithosphere

Repeat A, B, C & A

Many Dike-Flow Cycles

Fluid Dike
density ρ_f

Solid Dike
density ρ_d

Volcanic Infill
density ρ_i

Compensating
Layer, density ρ_s

Figure 3.

Figure 4.

Figure 5.

Figure 6.

Figure 7.