1 **An Integrated Kinematic and Geochemical Model to Determine Lithospheric**

2 **Extension and Mantle Temperature from Syn-Rift Volcanic Compositions**

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14 **Abstract**

15 We present an integrated kinematic and geochemical model that determines the

16 composition of melts and their residual source rocks generated by decompression

17 melting of the mantle during continental rifting. Our approach is to construct a unified

18 numerical solution that merges an established lithospheric stretching model which

19 determines the rate and depth at which melting occurs, with several compositional

20 parameterisations of mantle melting to predict the composition of primary melts. We

21 also incorporate a parameterisation for the rare earth elements. Using our approach,

22 we are able to track the composition of the melt fractions and mantle residues as

23 melting progresses. Our unified model shows that primary melt composition is

24   sensitive to rift duration and mantle temperature, with rapid rifting and higher mantle

25   temperatures producing larger melt fractions, at a greater mean pressure of melting,

26   than slower/cooler rifting. Comparison of the model results with primitive basalts

27   recovered from oceanic spreading ridges and rifted margins in the North Atlantic

28   indicates that rift duration and synrift mantle temperature can be inferred

29   independently from the appropriate geochemical data.

1. Introduction

Melting of the mantle occurs at rifted continental margins in response to decompression as mantle rocks are brought toward the surface by lithospheric extension. This melting is controlled by several parameters including: the initial mantle temperature and source composition; the rate of continental extension; the final degree of lithospheric stretching (β_{max}); and the initial lithospheric thickness [1]. Several studies have attempted to estimate some of these parameters from estimates of rift-related magmatic volume (e.g., [2],[3],[4],[5],[6]). This volume is usually estimated from wide-angle seismic data. Igneous products, added to the lower crust (i.e., underplated or intruded complexes) are identified as regions of high seismic velocity [2] relative to normal continental crust (e.g., [7]). Extrusive volcanics (i.e., flood basalts) are identified as high amplitude seaward dipping reflector sequences on the continent-ocean transition. Examples of both intrusive and extrusive magmatic rocks are found on "volcanic" margins such as the East Greenland margin [5],[8] the Faeroes margin and Hatton Bank [9]. In contrast, such features are absent at "non-volcanic" rifted margins, and the presence at some of these margins of exhumed serpentinized mantle, well-documented by both geophysical studies and by direct sampling in the southern Iberia Abyssal Plain (e.g., [10],[11]), shows that melting can be suppressed by different syn-rift conditions.

In both types of rifted continental margin, there are difficulties in measuring accurately the volume of magmatic products using seismic reflection techniques alone. This is due to the high impedance contrast between magmatic intrusions and the surrounding lower continental crustal material, as well as limited spatial resolution, which makes it difficult to image small and discontinuous bodies of

55 intruded material. However, it is possible to put constraints on the total intrusive

56 volume from the average lower-crustal velocity if the end member velocities of the

57 pre-existing crust and the intrusive rocks are known [9]. More importantly, the

58 volume of magmatic products is not constrained uniquely by the tectonic process

59 alone; rather it is also subject to variations in mantle temperature, composition and

60 strain rate.

61 An alternative approach to determining the rifting conditions (encompassing tectonic

62 geometry, strain rate, mantle temperature and composition) is to use the composition

63 of the volcanic products and, where available, upper mantle residues. [12] showed,

64 through modelling, how the rare earth element compositions of syn-rift volcanics on

65 the Labrador margin of eastern Canada and the North Sea Rift are sensitive to mantle

66 temperature, the degree of stretching and the duration of rifting. Using a similar

67 principle, we have developed an integrated kinematic and geochemical model to

68 determine the geochemical compositions of syn-rift melts and their residual mantle

69 source rocks under a variety of rifting conditions. From this model, we demonstrate

70 that the major, minor and rare-earth element geochemistry of the syn-rift melts and

71 their mantle residues are sensitive independently to the extension history and mantle

72 temperature. By inverting the approach, we compare the compositions of syn-rift

73 volcanic products from around the North Atlantic margin with those predicted by our

74 model and hence to infer the most probable rifting conditions prevalent during their

75 genesis.

76 **2. Rifted Margin Melting Model**

77 Various tectonic models for lithospheric stretching at continental margins have been

78 proposed including pure shear models [13], simple shear models [14], and models

79   involving depth-dependent stretching in different forms (e.g., [15]). The simplest

80   models to account for the main features of most rifted margins assume pure shear at a

81   uniform and finite rate. We employ one of these variants, the one-dimensional

82   lithospheric stretching model of [1], which determines several key parameters during

83   rifting: the temperature distribution throughout the lithosphere and asthenosphere and

84   the quantity of melt generated at each increment of rifting and melting.

85   To determine the temperature distribution during rifting, we solve the advection-

86   diffusion equation [16]:

87    $$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} - v\left(\frac{\partial T}{\partial z} + h\right),$$

88   where T is temperature, κ is the thermal diffusivity of the mantle, z is the vertical

89   coordinate, v is the vertical velocity of the mantle, and h is the adiabatic temperature

90   gradient. The temperature is corrected for the loss of latent heat of fusion using a

91   method similar to that of [17], assuming a value for the entropy of fusion of 400 J kg^{-1}

92   K^{-1}, and invoking mantle melting under dry solidus conditions. The advection of heat

93   with the melt and the latent heat released by solidification of the melt are ignored

94   since this heat is released in the crust, not in the mantle source region, and its effect on

95   melt generation is negligible [1]. Our model is constructed from a series of depth

96   regions, or cells, the thicknesses of which decrease as a function of rifting progress.

97   Throughout rifting, we assume a constant strain rate, parameterised by the rift

98   duration and the final degree of extension (β_{max}). We use a lithosphere with initial

99   thickness of 125 km that is in thermal equilibrium with the asthenosphere at the start

100   of rifting [18],[19]. Asthenospheric temperature is assumed to be invariant during

101   rifting. Other model parameters are as in Table 1 of [1].

102   The approach of [1] provides an ideal basis for modelling the melt product

103   geochemistry: first, it allows us to test all the rifting parameters of interest while

104   keeping the model simple; second, the melt chemistry parameterisations require

105   calculations of the degree of mantle depletion, temperature and pressure, all of which

106   are tracked through time; third, the composition of the melt and the residue can be

107   determined independently for each depth interval and at each time step in the model.

108   To calculate compositions we assume fractional melting, where the liquid phase

109   separates from the solid residue immediately after it has formed, followed by

110   extraction and lithospheric pooling without further reaction, equilibration or

111   refertilisation of the surrounding mantle matrix. While this assumption is probably a

112   simplification, recent studies of mantle melting and melt extraction based on ^{238}U–

113   ^{230}Th disequilibria find that incremental batch melting and rapid melt extraction is a

114   good approximation of the actual melting process in sub-oceanic mantle [20].

115   Similarly U–Th–Pa–Ra studies of oceanic basalts [21] find results that are consistent

116   with rapid melt migration and extraction via high-porosity channels, with minimal

117   melt-matrix interaction, as a general means of melt extraction from the upper mantle.

118   In addition, the parameterisations we use here enable sequential integrated melting

119   across phase boundaries in the mantle, accommodating melting in both the garnet and

120   spinel lherzolite fields (e.g. [22]). These simplifications allow us to implement several

121   melting schemes and parameterisations in our kinematic and thermal model for

122   comparison.

123   **3. Major Element Composition**

124   Three different major element parameterisations have been implemented [17], [23],

125   [24]. The first two use similar methods to determine melt composition from

126   experimentally derived partition coefficients. In contrast, [17] use empirical fits to

127   data from laboratory melting experiments to derive functions, thus describing

128   effective bulk partition coefficients of each element and the source mantle mineralogy

129   as melting proceeds.

130   **3.1 Source Composition**

131   Before considering the compositions of the magmatic product predicted by our

132   modelling, it is important to note that the composition of syn-rift melts are particularly

133   dependent on the composition of the upper-mantle source. In our modelling, we use a

134   selection of upper-mantle sources, chosen from a variety of studies of tholeiite

135   sources, to both assess the effects of source composition on our results and to inform

136   our choice of a source that yields melts that most closely resemble syn-rift volcanic

137   products.

138   In their modelling, [17] use a 'Depleted Mantle' source (DM) made by removing 40%

139   of the mass of a primitive mantle composition [25] by the production of average

140   continental crust (after [26]). [23] use three different sources: (i) 'Hawaiian Pyrolite'

141   is a synthetic, enriched and fertile spinel lherzolite used in experimental studies of the

142   genesis of Hawaiian tholeiite (after [27]), that is consequently more fertile (i.e., cpx-

143   rich) than 'DM'; (ii) 'Tinaquillo lherzolite' [28], a naturally occurring depleted or

144   refractory spinel lherzolite that is generally more depleted in incompatible elements

145   (except Al and Ca) than DM. 'Tinaquillo lherzolite' can be produced by invoking

146   ~5% more melting of a primitive mantle composition than that which has been

147   invoked to produce 'DM'; and (iii) MORB pyrolite 90, a synthetic MORB source

148   calculated by adding 24% of a mixture comprising 83% N-MORB and 17% primitive

149   olivine, to 76% harzburgite [29]. The [24] source is a naturally occurring depleted

150 abyssal peridotite that is close to the MORB pyrolite 90 composition. Unlike the other

151 sources mentioned here, it is a real rock, recovered from oceanic fracture zones and as

152 such, it is attractive since it represents a relatively fertile upper-mantle composition.

3.2 Niu and Batiza [23] and Niu [24] major element parameterisation

154 The composition of the accumulated melt fraction (C^L) is defined by [23] as the

155 integral:

$$156 \quad C^L = \frac{1}{X}\int_0^X \frac{C^R}{D+X(1-D)}dX,$$

157 where X is the total melt fraction, C^R is the initial composition of the residue or

158 'solid' fraction, and D is the partition coefficient. This integral is evaluated at each

159 time step (i) using:

$$160 \quad C_i^L = \frac{X_{i-1}C_{i-1}^L + X_i - X_{i-1}C_i^{batch}}{X_i},$$

161 where the composition of the incremental batch melt (C_i^{batch}) is calculated from the

162 residue composition at the end of the previous time step using:

$$163 \quad C_i^{batch} = \frac{C_{i-1}^R}{D+(X_i - X_{i-1})(1-D)}.$$

164 It follows that, for the first increment of melting, $C^L = C^{batch}$ and the residue

165 composition C_{i-1}^R is that of the source rock. The residue composition is recalculated at

166 end of each time step from the source rock composition (C_0^R) using:

$$167 \quad C_i^R = \frac{C_0^R - X_i C_i^L}{1-X_i}.$$

8

168   Partition coefficients are determined as a function of melt fraction and, with the

169   exception of TiO$_2$, of pressure (P). [23] determine the partition coefficient of each

170   major element completely independently using:

171    $D = e + fX + \dfrac{g}{X} + hP + i\dfrac{P}{X}.$

172   The constants e, f, g, h, and i depend on the source composition.

173   The parameterisation of [24] takes the same form as that of [23] for all partition

174   coefficients with the exception of D_{FeO}. Niu (1997) relates D_{FeO} to D_{MgO} using the

175   following equation:

176    $D_{FeO} = 0.317 + 0.0037X - 0.00346P + 0.213D_{MgO}.$

177   For low accumulated melt fractions (<10%) the melt compositions determined by

178   both [23] and [24] are incorrect for some elements. The problem is particularly

179   obvious for the Hawaiian pyrolite and Tinaquillo lherzolite source compositions. For

180   these source compositions, as the melt fraction tends toward zero the partition

181   coefficient for FeO falls to zero and for CaO tends toward very high values. This

182   results in a high degree of instability in the models, with instabilities propagating

183   through the calculated compositions of melt and residue from one increment of

184   melting to the next. To reduce this problem, [23] provide an alternative set of partition

185   coefficients for CaO and FeO which give stable results for the Hawaiian pyrolite and

186   Tinaquillo lherzolite sources down to accumulated melt fractions of 5%. We use these

187   alternative partition coefficients to calculate the melt and source residue compositions

188   for accumulated melt fractions down to 5%. For each increment of melting up to an

189   accumulated melt fraction of 5%, we assume that the partition coefficients are fixed.

190 In this way, as the accumulated melt fraction passes 5%, both the melt and residue

191 compositions match those that would have been formed if the first melt increment had

192 been at 5%. The partition coefficients for the source compositions 'MORB pyrolite

193 90' [23],[24] appear to yield a stable set of compositional solutions, even at very low

194 accumulated melt fractions (<1%). For these source compositions we follow the same

195 procedure as above but fix the minimum partition coefficient to that calculated for an

196 accumulated melt fraction of 1%.

197 **3.3 Watson and McKenzie [17] major element parameterisation**

198 [17] determine major element melt composition as a function of pressure and melt

199 fraction using the method of [30] but with modified parameters. Their

200 parameterisation gives instantaneous melt composition at each stage of melting but,

201 while it accounts for the mineralogical depletion of the source, the residue

202 composition is not explicitly defined and we do not calculate it. The instantaneous

203 melt composition (c) was initially parameterised following [30]:

204 $c = a + b(1 - X)^{(1-D)/D}$.

205 To improve the fit to the data, however, [17] sub-divided the melting range into three

206 phase regions (plagioclase-, spinel- and garnet-lherzolite respectively), $0 < X < X_1$,

207 $X_1 < X < X_2$ and $X_2 < X < 1$, and provide separate functions for each. They relate

208 the composition of the accumulated melt fraction to the instantaneous melt

209 composition using:

210 $c = \dfrac{d}{dX}(XC^L)$,

211 then C^L is given by:

212 $\quad C^L = a + \dfrac{bD}{X}\left[1 - (1 - X)^{1/D}\right] + \dfrac{d}{X}.$

4. Rare Earth Element Composition

214 The REE composition is determined using the parameterisation of [25]. [25] depleted

215 earth source composition is given in Table 2. The composition of the melt and residue

216 is determined using the same equations as for the major elements in [23] and [24].

217 The partition coefficients are dependent on the plagioclase, spinel and garnet stability

218 fields. [25] provide the proportion of each mineral present in each stability field, and

219 the partition coefficient for each REE within each mineral. For each REE the partition

220 coefficient for the whole mineral assemblage is the sum of each mineral's partition

221 coefficient multiplied by the proportion of that mineral. The stability fields are

222 defined as 25 km and 35 km for spinel-in and plagioclase-out respectively; values that

223 [25] consider reasonable estimates based on the experimental results of [31]. Garnet-

224 in and spinel-out limits are defined by pressure as a function of temperature; the

225 spinel-out boundary is given by:

226 $\quad P = \dfrac{T + 400}{666.7},$

227 and the garnet-in boundary is given by:

228 $\quad P = \dfrac{T + 533}{666.7}.$

229 The transition between regimes is assumed to be linear.

5. Model Results

We have run our model for a range of mantle potential temperatures (1200-1500ºC) to a stretching factor $\beta_{max} = 50$, which is found to simulate infinite stretching and hence steady-state melt production at a mid-ocean ridge. To explore the effect of strain rate on melt volume and composition, we have, for each temperature step, run the model for a range of rift durations of 1-50 m.y. Results for long rift durations describe behaviour at low strain rates, while results for short durations describe behaviour at high strain rates.

5.1 Melt volume, melt fraction and rifting history

As rifting rates slow, an increasingly significant amount of heat is lost from the upper mantle by conduction through the lithosphere. This heat loss suppresses shallow mantle melting, resulting in a reduced melt fraction and a lower total melt volume or thickness (Fig. 1). Because the base of the melting region is insensitive to rift duration, as rifting rates decrease, the mantle continues to provide melts during rifting, weighting the total aggregated melt composition towards deeper melts and hence effectively increasing the mean pressure of melting (Fig. 2a). A similar result is also seen with lower β_{max} (i.e., with decreasing strain). Aggregated melt compositions approach steady state only after stretching by a factor of $\beta_{max} \geq 5$ (Fig. 2b). With higher mantle temperatures the base of the melting zone moves deeper [30] while the shallowest depths of melting remain unaffected, weighting the aggregated melt composition towards higher mean pressures. The dependence of melting on rifting rate, however, is the same regardless of mantle temperature. Long duration rifting suppresses shallow mantle melting, causing an increase in the mean pressure of melting, a reduction in the mean melt fraction and a lower melt volume or thickness.

254   As a result, the mean melt fraction, total melt volume and mean pressure of melting

255   are critically dependent on both the rifting rate and the initial mantle temperature (Fig.

256   2c,d).

257   Variations in the extent and depth of mantle melting underlie the changes in melt

258   composition (discussed below). Thus it also follows that the composition of syn-rift

259   melts are themselves indicators of the rifting conditions under which they were

260   generated. In the following sections we explore the compositional effects predicted by

261   our modelling and compare actual syn-rift magmatic products to those results in an

262   attempt to distinguish between various rifting conditions.

263   **5.2 Major element trends with varying temperature and rift duration**

264   Despite the effects of using different mantle source compositions, all the

265   parameterisations used in our model yield similar chemical trends for the composition

266   of melts produced under varying mantle temperature and rift duration conditions

267   (Figs. 3, A1-A4). In general, those elements showing decreasing concentrations with

268   increasing mantle temperature (Na, Ti, K, Al and Ca), also show decreasing

269   concentrations with decreasing rift duration (i.e., increasing strain rate). This result

270   reflects changes in the mean melt fraction and subsequent melt volume (Fig. 2c,d) and

271   is to be expected for elements that are generally incompatible in bulk mantle

272   compositions (i.e., such that their concentration in the melt decreases as the extent of

273   melting increases).

274   Two elements (Si and Fe), however, do not show this simple relationship. For the

275   model results using the parameterisations of [23] (Figs. A1-A3) and [24] (Fig. A4), Si

276   decreases in concentration with both increasing mantle temperature and rift duration

277   (i.e., as strain rate decreases). Fe shows an opposite effect, increasing with hotter

278   mantle temperature and longer rift duration (i.e., as strain rate decreases). For the

279   model results using the parameterisation of [17], Si concentrations generally decrease

280   with increasing mantle temperature and rift duration while Fe increases with mantle

281   temperature but is unaffected by rift duration (Fig. 3). The behaviour of Si and Fe is

282   governed by the competing effects of mean melt fraction and mean pressure of

283   melting. Both Si and Fe are particularly sensitive to pressure of melting [32], [33],

284   such that the bulk partition coefficient of silica increases at higher pressures (i.e.,

285   silica becomes more compatible at greater depth), while that for iron decreases. The

286   net result is that while rift duration increases, the mean melt fraction decreases and the

287   mean pressure of melting increases, resulting in lower silica and higher iron

288   concentrations.

289   Our model results show that, as a result of the differing degrees of sensitivity of the

290   major elements to mean melt fraction and mean pressures of melting, any point within

291   mantle-temperature versus rift-duration space yields a unique melt composition.

292   While these unique melt compositions vary depending on the parameterisations and

293   source compositions used in the model, their uniqueness means that the compositions

294   of real rocks may be used as a proxy for mantle temperature and strain rates during

295   continental rifting. While our model is designed to predict the composition of melts

296   generated during continental rifting, it should also predict the composition of N-

297   MORB if rifting is effectively instantaneous and the mantle temperature lies close to

298   that believed to be applicable to the mid-ocean ridge region.

299   **5.3 Residue compositions**

300   The parameterisation of [23] allows the major element composition of the source

301   residue to be continuously recalculated and tracked as melting progresses. The final

302 composition of the residual mantle (Fig. 4) is that which has undergone the most

303 extensive melting by the end of the rifting period. Similar to the compositions of the

304 melt products, the composition of the mantle residue is affected by the pressure and

305 temperature history of melting, as well as by its starting composition. In general,

306 higher mantle temperatures and shorter rift durations result in greater concentrations

307 of elements compatible in the mantle (e.g., magnesium and iron) and decreasing

308 concentrations of incompatible elements (silica, potassium, sodium aluminium,

309 titanium and calcium). While the predicted mantle residue compositions allow the

310 potential for comparison with actual rocks, abyssal peridotites are rarely recovered

311 without significant serpentisation, which significantly affects the whole rock

312 composition by metasomatic exchange between the rock and seawater.

313 **5.4 REE compositions**

314 Incorporating the partition coefficients for the rare earth elements (REE) and the

315 parameterisation of [25] in our model allows for a calculation of the REE

316 concentrations in the aggregated melt products (Fig. 5). The REE show a general

317 increase in partition coefficient in mafic liquids with decreasing atomic mass –

318 although the bulk partition coefficients between whole rock and melt, in mantle

319 mineralogy, remain substantially less than one. Hence the behaviour of the REE, with

320 varying mantle temperature and rift duration, are similar to that of the incompatible

321 major elements (e.g., sodium, potassium and titanium). As the mantle temperature

322 increases and the rift duration decreases, the concentrations of the REE decrease

323 rapidly. An exception to this behaviour is hafnium, which decreases in concentration

324 sharply between low and moderate mantle temperatures and at short rift durations, but

325 changes little at higher temperatures and at longer rift durations. The reason for this is

326 the pressure dependant behaviour of hafnium such that is compatible in garnet [34].

327   At high mantle temperatures and slow rifting, the mean pressure of melting is high.

328   This effect weights the aggregated melt composition towards those increments drawn

329   from the garnet-lherzolite stability field where hafnium is retained in the residual

330   garnet.

331   **6. Comparison with actual rocks**

332   In the following section, we compare the results of our model against actual volcanic

333   products recovered from rifted margins. Like most mantle melting models, our model

334   predicts the composition of primary melts, generated under a variety of conditions.

335   These melts undergo variable degrees of fractional crystallisation before being

336   erupted [35], [36], the effects of which are to increase the concentrations of

337   incompatible elements and reduce the concentrations of compatible elements. The

338   potential consequences of these effects, when comparing actual volcanic products

339   with our model results, are to overestimate the rift duration and underestimate the

340   mantle temperature. In order to minimize these effects we have chosen to compare

341   only the most primitive rocks (i.e. those from a narrow range of high MgO

342   concentrations), allowing us to discriminate between different rifting durations and

343   initial mantle temperatures.

344   The results of these comparisons are shown in two ways: as a multi-element spider

345   diagram (where calculated elemental concentrations are normalized to a known

346   reference material, such as N-MORB), and as misfit diagrams (which express the

347   extent of misfit between a particular whole rock composition and a range of

348   compositions, each calculated for specific mantle temperatures and rift durations).

349   The misfit (f_m) for each unique combination of mantle temperature and rift duration

350   is calculated from the following equation:

351     $$f_m = \frac{1}{N} \sum_{i=1}^{N} \frac{(E_{ip} - E_{io})^2}{(\delta_{min}^{max} E_{ip})^2}$$

352   where E_{ip} is the predicted concentration for the ith major element, at a unique mantle

353   temperature and rift duration; E_{io} is the observed concentration for that element in the

354   reference rock; $\delta_{min}^{max} E_{ip}$ is the range of predicted concentrations for that element; and

355   N is the number of elements.

356   Hence the final misfit value for a particular rock is the average of the root-mean-

357   squares (rms) of the misfits of all the elements to the entire range of model results.

358   Each element's misfit is normalised to the range of that element's concentration as

359   predicted by the model, so that all elements, regardless of their percentage of the total

360   rock, are given equal weighting in the misfit. It should be noted that some element

361   concentrations are more accurately and precisely measured than others. Also, the

362   behaviour of some elements during melting is better known, while other elements are

363   more or less susceptible to secondary alteration. However, weighting one element

364   more than another cannot be justified without knowing the history of each element

365   during the melting, fractional crystallisation and alteration stages of a particular rock.

366   The only exception that we make is for potassium. Potassium is not included in the

367   misfit functions because its concentration is poorly known in the depleted upper

368   mantle (see below) – it is often associated with metasomatic mantle enrichment and it

369   is especially strongly affected by secondary alteration.

6.1 Comparison with N-MORB

370

371 As an initial assessment of our model, we compare normal mid-ocean ridge basalt (N-

372 MORB, [37]) to the predicted compositions of primary melts using various

373 parameterisations, mantle sources, mantle potential temperatures (1300°C and

374 1500°C) and rift durations of 1 and 50 m.y. The resultant liquids, normalized to N-

375 MORB, show the effects of increasing rift duration and mantle temperature (Fig. 6).

376 These effects are especially apparent for the incompatible elements (K, Ti, Na and Ca)

377 which increase in concentration for lower melt fractions and longer rift durations, and

378 which mimic the effects of more fertile mantle sources (e.g., Hawaiian pyrolite).

379 Of the five parameterisations and source compositions used in our model, the results

380 obtained by using the parameterisation of [17], at 1300°C and a rift duration of 1 m.y.,

381 produce a composition that most closely resembles N-MORB. Figure 7 shows results

382 for the REE concentrations, normalised to N-MORB [37], predicted for temperatures

383 of 1300°C and 1500°C and rift durations of 1 m.y. and 50 m.y. The resulting

384 predicted liquid produced at a mantle temperature of 1300°C and almost

385 instantaneous rifting (1 m.y.) compares favourably with N-MORB while those melts

386 produced at higher temperatures are more depleted, and that generated at 1300°C and

387 50 m.y. is more enriched.

388 Despite the close fit between the predicted major elemental concentrations and the

389 composition of N-MORB, potassium remains poorly reproduced. This problem exists

390 for all five parameterisations and sources and probably results from either incorrect

391 assumptions for the concentration of potassium in the source, poorly constrained

392 partition coefficients, or anomalously low potassium concentrations in our

393 normalising N-MORB composition. This observation reinforces our decision not to

394 use potassium concentrations in our misfit function.

395 Figures 8a and b shows the minimum misfit between primitive N-MORB [37] and the

396 model results incorporating the parameterizations of [17] and [24] using the MPY90

397 source. The misfit values are minimised at a temperature and rift duration that yields a

398 liquid that lies closest in composition to primitive N-MORB. Of these results, the

399 misfit generated by the model incorporating the parameterization of [17] is minimised

400 at a mantle potential temperature of 1260°C and a rift duration of 2 m.y. This is close

401 to the mantle potential temperature of 1280°C, used by [17] in their original

402 parameterization to model parental N-MORB genesis at a mid-ocean ridge. In

403 comparison, the other parameterizations used in our model generally yield poorer

404 misfits with minima at higher mantle temperatures and/or longer rifting duration (i.e.,

405 slower rifting rates). For example, using the parameterization of [24] yields higher

406 mantle temperatures and longer rift durations, respectively, of: 1430°C and 14 m.y.

407 for the MORB pyrolite source MPY90; 1406°C and 1 m.y. for the Hawaiian pyrolite

408 source; and 1420°C and 8 m.y. for the Tinaquillo lherzolite source.

409 **6.2 Comparison with North Atlantic Margin Volcanic products**

410 In the following section, we calculate the misfit functions for a number of volcanic

411 suites from around the northwest Atlantic margin, formed during the initial opening of

412 the North Atlantic. Several volcanic sequences have been collected at each site by

413 DSDP and ODP drilling expeditions. In comparing these rocks with our model, we

414 have chosen an average of the most primitive (i.e., highest MgO - at least 7 wt.%),

415 least phyric, and least altered rocks (e.g., having the lowest reported 'loss on ignition'

416 values) of basaltic composition from each sequence. The geotectonic setting of these

417 break-up lavas varies considerably: from the volcanic rifted margin of East Greenland

418 at 63°N (ODP Sites 915, 917, 918, 989, 990) and its conjugate at Edoras Bank margin

419 (DSDP Site 553 and 554), to the 'non-volcanic' margin of Goban Spur (DSDP Sites

420 550 and 551). These rifts were initiated at varying distance and time from the mantle

421 melting anomaly (Icelandic plume) that was later to form Iceland. Hence we might

422 predict a variation in mantle temperature during rifting with relatively higher mantle

423 temperatures closer to the centre of the melting anomaly and lower temperatures both

424 further away and prior to the start of the anomaly [38].

425 We have chosen to calculate the major element misfits to our model using the

426 parameterization of [17], since this method gives the best fit to N-MORB.

427 Furthermore, for the remaining parameterizations other than [17], most of the syn-rift

428 volcanic suites plot out of the range of the predicted mantle temperature and rift

429 duration. Because we have not corrected for AFC processes, the minimum misfits

430 predict relative mantle temperatures and rifting rates rather than absolute conditions

431 during rifting. Not all of the whole rock analyses available for comparison to our

432 model include a full set of REE concentrations, so we have decided to omit the REEs

433 from any of the misfit functions. Figure 5 shows that the REEs are more sensitive to

434 mantle temperature than rift duration. However, since rift duration is the main

435 parameter affecting mean pressure of melting, we would recommend including the

436 concentration of heavy REEs (e.g., Lu and Hf) as well as the light REEs (e.g., La and

437 Sm) in any future misfit functions.

438 **6.2.1 East Greenland margin**

439 ODP Legs 152 and 163 drilled a transect across the East Greenland volcanic margin at

440 63ºN, where drilling penetrated a thick succession of seaward dipping reflector

441 sequence (SDRS) volcanic rocks that erupted during break-up of the North Atlantic in

442 the early Tertiary. The western-most holes are well within the influence of the proto-

443 Iceland plume [38], while the easternmost holes are on thickened oceanic basement.

444 With both a decrease in age and increase in distance east, the SDRS lava succession is

445 thought to represent a transition from continental break-up to seafloor spreading [9],

446 [39], [40]. On the basis of their position and age, the lava sequences recovered in the

447 holes can be ordered according the progression from continental rifting to initial

448 seafloor spreading: 917A Lower lavas are the oldest and were erupted during

449 continental rift initiation with Site 917A Upper and Site 989 lavas being later syn-rift

450 eruptions on stretched continental crust. Site 918D lavas were erupted at the transition

451 from the late continental rifting to early oceanic spreading. Holes 990A and 915A are

452 both located close together and recovered SDRS lavas forming the youngest

453 extrusives prior to the onset of full seafloor spreading.

454 Recent seismic models ([8], [41]) suggest a crustal thickness along the drilling

455 transect of up to 18 km [41] at the time of break-up, reducing to 8-10 km thickness 6-

456 12 m.y. after break-up [8]. [8] attribute the initial melt thickness to a transient elevated

457 temperature anomaly of ~1400±50°C in the asthenosphere; while transient active

458 asthenospheric upwelling could produce the same melt thickness pattern, the location

459 is thought to be sufficiently distal to the Iceland hotspot that this process is not a

460 significant factor.

461 The parameters in our melting model do not incorporate the effect of a transient

462 temperature anomaly. Such a transient could be achieved by adding a high

463 temperature layer directly beneath the continental lithosphere [41], which travels

464 upward, and passes through the melting region, as rifting proceeds. [42] show that a

465   1425-1525°C, 50 km thick layer can produce a similar melt thickness pattern to the

466   seismic observations [8]. In such models, the majority of the melt, especially that

467   located near to the ODP Sites where initial break-up occurred, is formed at the

468   elevated asthenospheric temperature. In this case, our melting model calculates the

469   average melt chemistry formed during the entire rift period, and is therefore biased

470   toward the chemical signature of the high temperature melt.

471   Figure 8c shows the misfit functions for the East Greenland margin early syn-rift

472   SDRS lavas (Sites 917L and 917U, 989) while Figure 8d shows data from the SDRS

473   for the later and/or more oceanward syn-rift sequence. In each case the mantle

474   temperature prediction is more constrained than the rifting rate, with both groups of

475   SDRS lavas predicted to have formed under similar conditions: slow rifting rates (i.e.,

476   rift duration of ~40 m.y.) and with mantle temperatures of ~1400°C. The elevated

477   mantle temperatures are consistent with the presence of a thermal anomaly associated

478   with the Icelandic mantle plume [2], [9], [38], [43]. An excess temperature of ~140°C

479   for the syn-rift mantle beneath East Greenland, compared with asthenosphere yielding

480   N-MORB, is also compatible with the enhanced thickness of SDRS lavas and the high

481   seismic velocities and thicknesses of the magmatic lower-crustal intrusion that is

482   characteristic of this margin [5], [9], [40]. As seafloor spreading progressed, the

483   mantle temperature decreased by ~70–80°C over the first 10 Ma following continental

484   breakup [40]. This decrease might be a result of exhaustion of a finite hot layer of

485   mantle [42], a reduction in the diameter of the plume head (as proposed by [38]), drift

486   of the plume stem away from the region, a secular decline in the plume mantle

487   temperature, or a combination of all of these effects [4].

488   **6.2.2 Edoras Bank margin**

489   DSDP Leg 81 drilled three sites across the Edoras Bank on the western Atlantic

490   margin at 56ºN. Drilling at the sites penetrated SDRS lavas of late Paleocene/early

491   Eocene [44]. Figure 8e shows the average misfit functions for the Edoras Bank

492   margin basalts from holes 553 and 554. Sites 552, 553 and 554 are located

493   progressively westward (i.e., oceanward) across the rifted margin. Sites 553 and 554

494   recovered basalt that is sufficiently fresh to be used in our modelling, but Site 552

495   lavas are too strongly altered and contaminated by continental crust to be included in

496   our analysis. Site 553 is situated on stretched continental crust while Site 554 is on the

497   outer high between seaward dipping reflectors [4], [44], [45].

498   The average misfit functions of the remaining rocks constrain better the mantle

499   temperature than the rifting rate. The Edoras Bank margin appears to have started

500   rifting with a mantle that was between 1350 and 1375°C (i.e., 25-50°C cooler than the

501   ODP transect off East Greenland). This slightly lower mantle temperature is

502   consistent with the Edoras Bank margin being located (at 56°N) further to the south of

503   the East Greenland sites at 63°N, and hence at a greater distance from the centre of

504   the Icelandic mantle plume. It is also consistent with slightly thinner SDRS and the

505   thickness of lower-crustal magmatic intrusions at the Edoras Bank margin (at 56°N),

506   compared with the East Greenland margin at 63°N [4]. The subsidence calculations in

507   [4] show a somewhat higher mantle temperature than our estimates from the

508   geochemistry, but the overall trends are equivalent, i.e., the temperature decreases

509   away from the plume centre. We also note that the subsidence calculations assume a

510   single phase of passive upwelling. If there were a component of active upwelling in

511   the mantle beneath the initial rift at the time of continental break-up, as indeed may be

512   the case [41], rather lower temperatures would be required to give the same

513   subsidence. If this were the case then the subsidence and geochemical mantle

514   temperature constraints could be very similar.

515   **6.2.3 Goban Spur margin**

516   DSDP Leg 80 drilled several sites on the western Atlantic margin at 49ºN. Unlike the

517   East Greenland and Edoras Bank margins, which are considered to have been

518   influenced by elevated mantle temperatures, the Goban Spur margin does not have a

519   significant high velocity lower-crustal section and is considered have had little syn-rift

520   magmatism [46]. Drilling recovered basalts at two sites, 550 and 551, both on the

521   ocean-continent transition. At these locations, the structure of the Goban Spur margin

522   is complex, with an east-west oriented offset forming a ridge that separates Site 551

523   (to the northeast) from Site 550 (to the southwest). Wide-angle seismic data indicate

524   that Site 551 is underlain by highly thinned continental crust [47]. There are no wide-

525   angle seismic constraints for Site 550, though this site lies about 50 km along-strike

526   from a region where the basement is interpreted as serpentinised mantle [47]. If this

527   basement is exhumed continental mantle, as inferred on the west Iberia margin [11], it

528   is then unclear which site hosts lavas formed later in the rifting process. Using lavas

529   from both sites 550 and 551 to predict the rifting conditions for this margin, Figure 8f

530   shows the average misfit functions that have a minimum for a mantle temperature of

531   1260-1285°C and a rift duration of 8-13 m.y. The inferred rift durations from melt

532   compositions are close to the estimated duration of 14-22 m.y. based on stratigraphic

533   evidence [47], while melting temperatures are close to those for N-MORB genesis.

534   Our modelling suggests that syn-rift mantle temperatures for Goban Spur were ~80°C

535   cooler than the East Greenland margin at 63°N, and 60°C cooler than the Edoras Bank

536   margin. This result is not especially surprising, since the mid- to late Cretaceous age

537   of rifting at Goban Spur was prior to the known initiation of the Icelandic mantle

538   plume [43].

539   **7. Summary and conclusions**

540   We have developed a numerical model that integrates kinematic and geochemical

541   parameters during continental stretching and rifting to allow us to predict the

542   composition of magmatic products generated under a range of continental rifting

543   conditions. Using a modular scheme, we have integrated the lithospheric stretching

544   model of [1], with three different major element parameterisations: [17], [23] and

545   [24]. We also incorporate a parameterisation for the rare earth elements based on [25].

546   Our model shows how the geochemistry of basalts erupted during rifting can be used

547   to infer relative differences in mantle temperature and rift duration. These predictions

548   can, in turn, be used to predict the volume of syn-rift magmatism, as expressed both

549   by extrusive lavas that form SDRS packages and intrusion of the stretched continental

550   lithosphere, and hence the thermal evolution and uplift/subsidence histories for rifted

551   continental margins. Comparison of our model results, using a range of

552   parameterisations, with actual mid-ocean ridge basalts (i.e., instantaneous rifting)

553   reveals close reproduction of N-MORB compositions when using the parameterisation

554   of [17]. Calculated misfit functions between predicted compositions and actual

555   volcanic rocks from around the North Atlantic margin (Fig. 8c-f) confirm that syn-rift

556   lavas erupted closer to the centre of the Icelandic hotspot were derived from hotter

557   average mantle temperatures than those erupted further away. The composition of

558   rocks erupted within the influence of the Iceland mantle plume indicate higher mantle

559   temperatures than those erupted on non-volcanic margins away from the influence of

560   mantle plumes.

561 Because our model predicts the composition of primitive melts that are in equilibrium

562 with the mantle, we recommend comparing only the freshest and most primitive

563 basaltic rocks with our model results (e.g., ideally high-magnesium glasses).

564 Fractional crystallisation will result in an apparent over estimate of rift duration

565 and/or an underestimate of mantle potential temperature. We also recommend that

566 care be taken to screen samples for evidence of crustal contamination, either through

567 isotopic analysis (Pb and Nd ratios) and/or scrutiny of the high-field strength trace

568 elements that are indicators of crustal contamination (e.g. Zr and Nb). We further

569 suggest that any misfit function utilising REE compositions should weigh equally the

570 light and heavy REE's to maximise sensitivity to both the mean extent and the

571 pressure of melting, which are proxies for mantle temperature and rift duration

572 respectively.

576 **Appendix**

577 Supplementary data associated with this article consists of the major element

578 composition of the melt fraction for varying rift duration and asthenospheric mantle

579 potential temperature for four additional source compositions (Figures A1-A4).

580

References

581 **References**

582 [1] J.W. Bown, R.S. White, The effect of finite extension rate on melt generation at

583 continental rifts, J. Geophys. Res. 100 (1995) 18011—18030.

584 [2] R. White, D. McKenzie, Magmatism at rift zones: the generation of volcanic

585 continental margins and flood basalts, J. Geophys. Res. 94 (1989) 7685—7729.

586 [3] C.E. Keen, R.C. Courtney, S.A. Dehler, M.-C. Williamson, Decompression

587 melting at rifted margins: comparison of model predictions with the distribution

588 of igneous rocks on the eastern Canadian margin, Earth Planet. Sci. Lett. 121

589 (1994) 403—416.

590 [4] A.J. Barton, R.S. White, Crustal structure of the Edoras Bank margin and

591 mantle thermal anomalies beneath the North Atlantic, J. Geophys. Res. 102

592 (1997) 3109—3129.

593 [5] J. Korenaga, W.S. Holbrook, G.M. Kent, P.B. Kelemen, R.S. Detrick, H.-C.

594 Larsen, J.R. Hopper, T. Dahl-Jensen, Crustal structure of the southeast

595 Greenland margin from joint refraction and reflection seismic tomography, J.

596 Geophys. Res. 105 (2000) 21591—21614.

597 [6] P.B. Kelemen, W.S. Holbrook, Origin of thick, high-velocity igneous crust

598 along the US East-Coast Margin, J. Geophys. Res. 100 (1995) 10077—10094.

599 [7] N.I. Christensen, W.D. Mooney, Seismic velocity structure and composition of

600 the continental crust: a global view, J. Geophys. Res. 100 (1995) 9761—9788.

601 [8] W.S. Holbrook, H.-C. Larsen, J. Korenaga, T. Dahl-Jensen, I. Reid, P. B.

602 Kelemen, J.R. Hopper, G.M. Kent, D. Lizzaralde, S. Bernstein, R.S. Detrick,

603   Mantle thermal structure and active upwelling during continental breakup in the

604   North Atlantic, Earth Planet. Sci. Lett. 190 (2001) 251—266.

605 [9] R.S. White, L.K. Smith, A.W. Roberts, P.A.F. Christie, N.J. Kusznir, Lower-

606   crustal intrusion on the North Atlantic continental margin, Nature 452 (2008)

607   460—464.

608 [10] S.M. Dean, T.A. Minshull, R.B. Whitmarsh, K.E. Louden, Deep structure of the

609   ocean-continent transition in the southern Iberia Abyssal Plain: II. The IAM-9

610   transect at 40º20'N, J. Geophys. Res. 105 (1999) 5859—5885.

611 [11] R.B. Whitmarsh, G. Manatschal, T.A. Minshull, Evolution of magma-poor

612   continental margins from rifting to seafloor spreading, Nature 413, 2001, 150—

613   154.

614 [12] M.-C. Williamson, R.C. Courtney, C.E. Keen, S.A. Dehler, The volume and

615   rare earth element concentrations of magmas generated during finite stretching

616   of the lithosphere, J. Petrol. 36 (1995) 1433—1453.

617 [13] D. McKenzie, Some remarks on the development of sedimentary basins, Earth

618   Planet. Sci. Lett. 40 (1978) 25—32.

619 [14] B. Wernicke, Uniform sense normal simple shear of the continental lithosphere,

620   Can. J. Earth Sci. 22 (1985) 108—125.

621 [15] M. Davis, N. Kusznir, Depth-dependent lithospheric stretching at rifted

622   continental margins, in: G.D. Karner (Ed.), Proceedings of NSF Rifted Margins

623   Theoretical Institute, Columbia University Press, 2004, pp. 92—136.

624 [16] W.H. Press, B.P. Flannery, S.A. Teukolsky, W.T. Vetterling, Numerical

625   Recipes in FORTRAN 77: The art of scientific computing, Cambridge

626   University Press, New York, 1992, 992 pp.

627 [17] S. Watson, D. McKenzie, Melt generation by plumes: a study of Hawaiian

628   volcanism, J. Petrol. 32 (1991) 501—537.

629 [18] B. Parsons, J.G. Sclater, An analysis of the variation of oceanfloor bathymetry

630   and heat flow with age, J. Geophys. Res. 82 (1977) 803—827.

631 [19] H.N. Pollack, D.S. Chapman, On the regional variations of heat flow,

632   geotherms, and lithospheric thickness, Tectonophysics 38 (1977) 279—296.

633 [20] M. Condomines, O. Sigmarsson, U–238–Th–230 disequilibria and mantle

634   melting processes: a discussion, Chemical Geology 162 (2000) 95—104.

635 [21] A. Stracke, B. Bourdon D. McKenzie, Melt extraction in the Earth's mantle:

636   Constraints from U–Th–Pa–Ra studies in oceanic basalts, Earth Planet Sci. Lett.

637   244 (2006) 97—112.

638 [22] A. Prinzhofera, C.J. Allègre, Residual peridotites and the mechanisms of partial

639   melting, Earth Planet Sci. Lett. 74 (1985) 251—265.

640 [23] Y. Niu, R. Batiza, An empirical method for calculating melt compositions

641   produced beneath mid-ocean ridges: application for axis and off-axis

642   (seamounts) melting, J. Geophys. Res. 96 (1991) 21753—21777.

643 [24] Y. Niu, Mantle melting and melt extraction processes beneath ocean ridges:

644   evidence from abyssal peridotites, J. Petrol. 38 (1997) 1047—1074.

645 [25] D. McKenzie, R.K. O'Nions, Partial melt distributions from inversion of rare

646   earth element concentrations, J. Petrol. 32 (1991) 1021—1091.

647 [26] S.R. Taylor, S.M. McLennon, The Continental Crust: Its Composition and

648   Evolution, Blackwell, Oxford, 1985, 312 pp.

649 [27] D.H. Green, Experimental melting studies on a model upper mantle composition

650 at high pressure under water-saturated and water-unsaturated conditions, Earth

651 Planet. Sci. Lett. 19 (1973) 37—53.

652 [28] A.L. Jaques, D.H. Green, Anhydrous partial melting of peridotite at 0-15 kb

653 pressure and the genesis of tholeiitic basalts, Contrib. Miner. Petrol. 73 (1980)

654 287—310.

655 [29] D.H. Green, W.O. Hibberson, A.L. Jaques, Petrogenesis of mid-ocean ridge

656 basalts, in: M.W. McElhinney (Ed.), The Earth: Its Origin, Structure and

657 Evolution, Academic Press, London, 1979, pp. 265—299.

658 [30] D. McKenzie, M.J. Bickle, The volume and composition of melt generated by

659 extension of the lithosphere, J. Petrol. 29 (1988) 625—679.

660 [31] D.C. Presnall, J.D. Hoover, High pressure phase equilibrium constraints on the

661 origin of mid-ocean ridge basalts, in: B. Mysen (Ed.), Magmatic Processes:

662 Physicochemical Principles, Geochem. Soc. Spec. Pub. 1, 1987, pp. 75—89.

663 [32] E.M. Klein, C.H. Langmuir, Global correlations of ocean ridge basalt chemistry

664 with axial depth and crustal thickness, J. Geophys. Res. 92 (1987) 8089—8115.

665 [33] C.H. Langmuir, E.M. Klein, T. Plank, Petrological systematics of mid-ocean

666 ridge basalts: Constraints on melt generation beneath ocean ridges, in: J. Phipps

667 Morgan, D.K. Blackman, J.M. Sinton (Eds.), Mantle Flow and Melt Generation

668 at Mid-Ocean Ridges, Geophys. Mon. 71, 1992, pp. 183—280.

669 [34] V.J.M. Salters, S.R. Hart, The hafnium paradox and the role of garnet in the

670 source of mid-ocean-ridge basalts, Nature 342 (1989) 420—422.

671 [35] R.J. Kinzler, T.L. Grove, Primary Magmas of Mid-Ocean Ridge Basalts 2.

672 Applications, J. Geophys. Res. 97 (1992) 6907—6926.

673   [36]   E. Stolper, A phase-diagram for mid-ocean ridge basalts: Preliminary results

674   and implications for petrogenesis, Contrib. Mineral. Petrol. 74 (1980) 13—27.

675   [37]   S.R. Hart, J. Blusztajn, H.J.B. Dick, P.S. Meyer, K. Muehlenbachs, The

676   fingerprint of seawater circulation in a 500-meter section of ocean crust

677   gabbros, Geochimica et Cosmochimica Acta 63 (1999) 4059—4080.

678   [38]   R.S. White, Rift-plume interaction in the North Atlantic, Phil. Trans. Roy. Soc.

679   A. 355 (1997) 319—339.

680   [39]   J.G. Fitton, A.D. Saunders, L.M. Larsen, B.S. Hardarson, M.J. Norry, Volcanic

681   rocks from the southeast Greenland margin at 63°N: Composition, petrogenesis,

682   and mantle sources, in: A.D. Saunders, H.C. Larsen, S.W. Wise Jr. (Eds.),

683   Proceedings of the Ocean Drilling Program, Scientific Results, vol. 152, Ocean

684   Drilling Program, College Station, Tex., 1998, pp. 331—350.

685   [40]   R.S. White, L.K. Smith, Crustal structure of the Hatton and the conjugate east

686   Greenland rifted volcanic continental margins, NE Atlantic, J. Geophys. Res., in

687   press.

688   [41]   J.R. Hopper, T.Dahl-Jensen, W.S. Holbrook, H.C.Larsen, D. Lizarralde, J.

689   Korenaga, G.M. Kent, P.B. Kelemen, Structure of the SE Greenland margin

690   from seismic reflection and refraction data: Implications for nascent spreading

691   center subsidence and asymmetric crustal accretion during North Atlantic

692   opening, J. Geophys. Res. 108 (2003) doi:10.1029/2002JB001996.

693   [42]   J.J. Armitage, T.J. Henstock, T.A. Minshull, J.R. Hopper, Modelling the

694   composition of melts formed during continental breakup of the Southeast

695   Greenland margin, Earth Planet. Sci. Lett. 269 (2008) 248—258.

696   [43] R.S. White, A hot-spot model for early Tertiary volcanism in the N. Atlantic, in:

697   A.C. Morton, L.M. Parson (Eds.), Early Tertiary Volcanism and the Opening of

698   the NE Atlantic, Geol. Soc. Spec. Pub. 39, 1988, pp. 3—13.

699   [44] D.G. Roberts, D. Schnitker, et al., Initial Reports of the Deep Sea Drilling

700   Project, vol. 81, U.S. Government Printing Office, Washington, 1984, 923 pp.

701   [45] J. Brodie, J.G. Fitton, Data report: Composition of basaltic lavas from the

702   seaward-dipping reflector sequence recovered during deep sea drilling project

703   leg 81 (Hatton Bank), in: A.D. Saunders, H.C. Larsen, S.W. Wise Jr. (Eds.),

704   Proceedings of the Ocean Drilling Program, Scientific Results, vol. 152, Ocean

705   Drilling Program, College Station, Tex., 1998, pp. 431—435.

706   [46] R.C. Maury, H. Bellon, H. Bougault, J.L. Joron, M. Bohn, P.C. de Graciansky,

707   Oceanic tholeiites from Leg 80 sites (Celtic Sea passive margin, northeastern

708   Atlantic): Geochemistry and mineralogy, in: P.C. de Graciansky, et al., Initial

709   Reports of the Deep Sea Drilling Project, vol. 80, U.S. Government Printing

710   Office, Washington, 1985, pp. 939—946.

711   [47] A.D. Bullock, T.A. Minshull, From continental extension to seafloor spreading:

712   crustal structure of the Goban Spur rifted margin, southwest of the UK,

713   Geophys. J. Int. 163 (2005) 527—546.

714

715 **Figure Captions**

716 *Figure 1:* (a) Upper panel: contoured plot of temperature (solid lines) versus depth,

717 through time, for the rift model with an asthenospheric mantle temperature of 1300°C,

718 a rift duration of 5 m.y., and β_{max} = 50. Equilibrium melt fraction is shaded yellow to

719 red and contoured with dotted lines. Lower panel: accumulated melt thickness through

720 time. (b) As (a) but for a rift duration of 20 m.y.

721 *Figure 2:* (a) Contoured average (mean) depth of melting at the end of the rift period;

722 the model has β_{max} = 50, an initial lithospheric thickness of 125 km, and a constant

723 strain rate throughout the rift period; a rift duration of 1 m.y. approximates

724 instantaneous rifting. (b) as (a) but for varying β and asthenospheric mantle

725 temperatures between 1250°C and 1500°C (black lines); the grey line represents the

726 depth at which maximum melting takes place. (c) Contoured melt fraction within the

727 region of melting and (d) melt thickness accumulated at the end of the rift period,

728 using the same model as (a).

729 *Figure 3:* The major element composition of the melt fraction in wt%, for varying rift

730 duration and asthenospheric mantle potential temperature, calculated using the

731 parameterisation of [17]. The model uses the same depleted mantle (DM) source

732 composition (spinel lherzolite) as employed by [30] and has an initial lithospheric

733 thickness of 125 km with a constant strain rate throughout the rift period, and β_{max} =

734 50.

735 *Figure 4:* Major element composition of the residue in wt%, for varying rift duration

736 and asthenospheric mantle potential temperature, calculated using the

737 parameterisation of [24]. The model assumes a Hawaiian pyrolite source composition

738 and has an initial lithospheric thickness of 125 km, a constant strain rate throughout

739 the rift period, and $\beta_{max} = 50$.

740 *Figure 5:* Rare earth element composition (ppm) of the melt fraction for varying rift

741 duration and asthenospheric mantle potential temperature, calculated using the

742 parameterisation of [25]. Model parameters are the same as in Figure 2a.

743 *Figure 6:* Multi-element spider diagram for the major element compositions of

744 aggregated melts, normalised to average North Atlantic N-MORB [37], calculated

745 from the five different parameterisations, for rift durations of 1 m.y. and 50 m.y. and

746 asthenospheric mantle potential temperatures of 1300°C and 1500°C. The plots show

747 that the parameterisations of [17], [23] and [24] (the latter two both using a MORB

748 pyrolite source) that invoke mantle temperatures of 1300°C produces aggregated

749 melts that are closest in major element composition to N-MORB.

750 *Figure 7:* Rare earth element compositions for melt fractions generated at

751 asthenospheric mantle potential temperatures of 1300°C and 1500°C, and rift

752 durations of 1 m.y. and 50 m.y., normalised to North Atlantic N-MORB [37].

753 *Figure 8:* Misfit diagrams between model outputs and measured compositions. (a)

754 The misfit between N-MORB (average northern Mid-Atlantic Ridge [37]) and our

755 model output using the parameterisation of [24] for a MORB pyrolite source, yields a

756 best fit mantle temperature of ~ 1430°C at 14 m.y. (b) The misfit between N-MORB

757 (average northern Mid-Atlantic Ridge [37]) and our model output using the

758 parameterisation of [17] yields a best fit mantle temperature of 1260°C and a rift

759 duration of 2 m.y. (c) Misfit function for the East Greenland margin early syn-rift

760 SDRS lavas (Sites 917L and 917U, 989). The minimum misfit values indicate where

761   the average of the composition of the lavas best fits the melt composition predicted by

762   our model using the parameterisation of [17]. (d) As (c) for later syn-rift eruptions at

763   sites 918D, 990A and 915A. (e) As (c) for the Edoras Bank margin syn-rift basalts. (f)

764   As (c) for the Goban Spur margin syn-rift basalts. The model indicates that mantle

765   temperatures beneath the Goban Spur margin were close to N-MORB conditions

766   during the Cretaceous rift to drift transition.

767

768   **Appendix**

769   ***Figure A1:*** The major element composition of the melt fraction in wt%, for varying

770   rift duration and asthenospheric mantle potential temperature, an initial lithospheric

771   thickness of 125 km with a constant strain rate throughout the rift period, and $\beta_{max} =$

772   50, calculated using the parameterisation of [23] for Hawaiian pyrolite source

773   composition.

774   ***Figure A2:*** The major element composition of the melt fraction in wt%, for varying

775   rift duration and asthenospheric mantle potential temperature, an initial lithospheric

776   thickness of 125 km with a constant strain rate throughout the rift period, and $\beta_{max} =$

777   50, calculated using the parameterisation of [23] for Tinaquillo lherzolite source

778   composition.

779   ***Figure A3:*** The major element composition of the melt fraction in wt%, for varying

780   rift duration and asthenospheric mantle potential temperature, an initial lithospheric

781   thickness of 125 km with a constant strain rate throughout the rift period, and $\beta_{max} =$

782   50, calculated using the parameterisation of [23] for MORB pyrolite 90 source

783   composition.

784   ***Figure A4:*** The major element composition of the melt fraction in wt%, for varying

785   rift duration and asthenospheric mantle potential temperature, an initial lithospheric

786   thickness of 125 km with a constant strain rate throughout the rift period, and $\beta_{max} =$

787   50, calculated using the parameterisation of [24].

788

(a) Contour plot with axes "rift duration (m.y.)" (horizontal, 10 to 50) and "mantle temperature (°C)" (vertical, 1200 to 1500). Contours labeled 30, 35, 40, 45, 50, 55, 60. Region labeled "no melting".

(b) Plot with axes "β_{max}" (horizontal, 1 to 50, logarithmic) and "mean depth of melting (km)" (vertical, 0 to 120). Curves labeled 1250°, 1300°, 1350°, 1400°, 1450°, 1500°.

(c) Contour plot with axes "rift duration (m.y.)" (horizontal, 10 to 50) and "mantle temperature (°C)" (vertical, 1200 to 1500). Contours labeled 0.15, 0.1, 0.05. Labeled "melt fraction".

(d) Contour plot with axes "rift duration (m.y.)" (horizontal, 10 to 50) and "mantle temperature (°C)" (vertical, 1200 to 1500). Contours labeled 20, 15, 10, 5. Labeled "melt thickness (km)".

Figure 6

Legend:
- + Watson & McKenzie (1991)
- ○ Hawaiian Pyrolite (Niu & Batiza, 1991)
- □ Tinaquillo Lherzolite (Niu & Batiza, 1991)
- △ MORB pyrolite 90 (Niu & Batiza, 1991)
- ☆ Niu (1997)

Figure 7
Click here to download Figure: fig7.eps
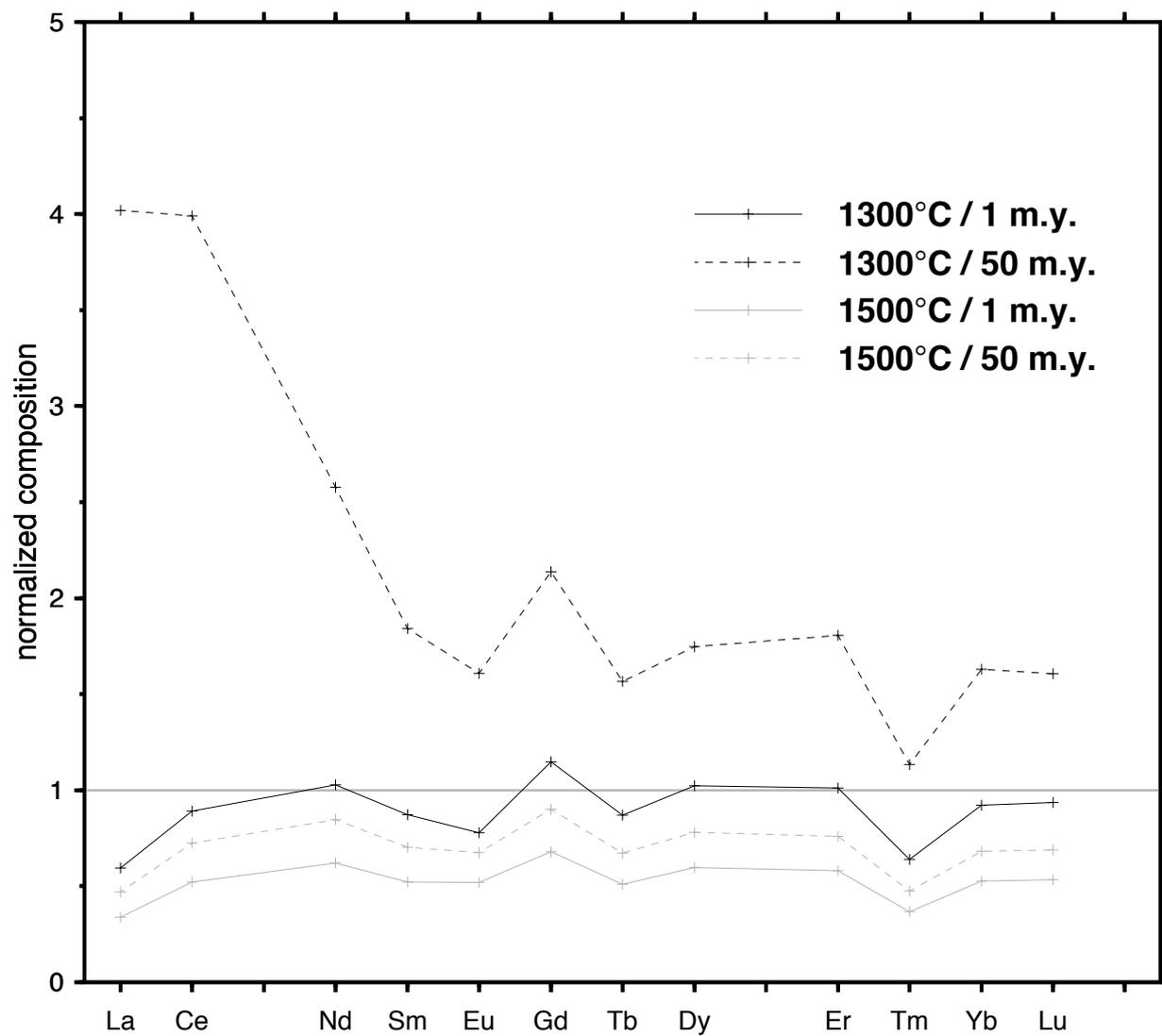
Figure 8

(a) Misfit to MAR N-MORB (Niu, 1997; MPY90)

(b) Misfit to MAR N-MORB (Watson & McKenzie, 1991)

(c) East Greenland (early syn-rift sites 989, 917)

(d) East Greenland (late syn-rift sites 990, 918 & 915)

(e) Edoras Bank

(f) Goban Spur
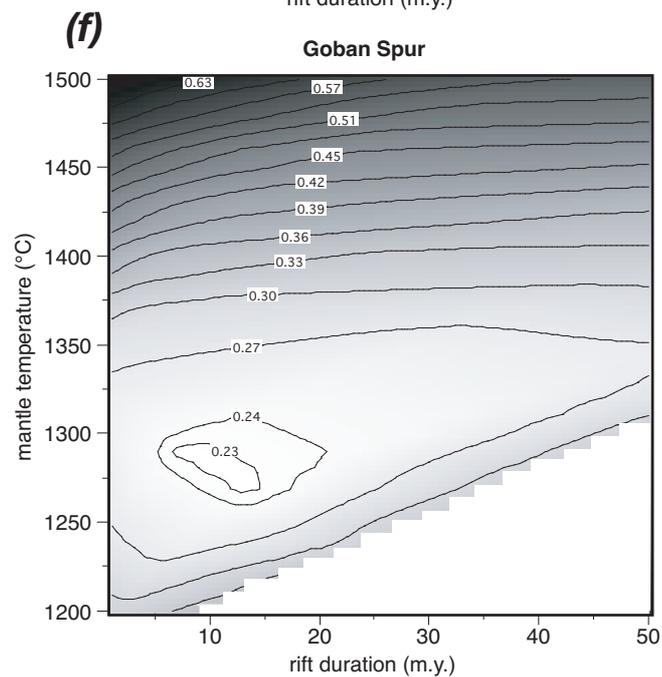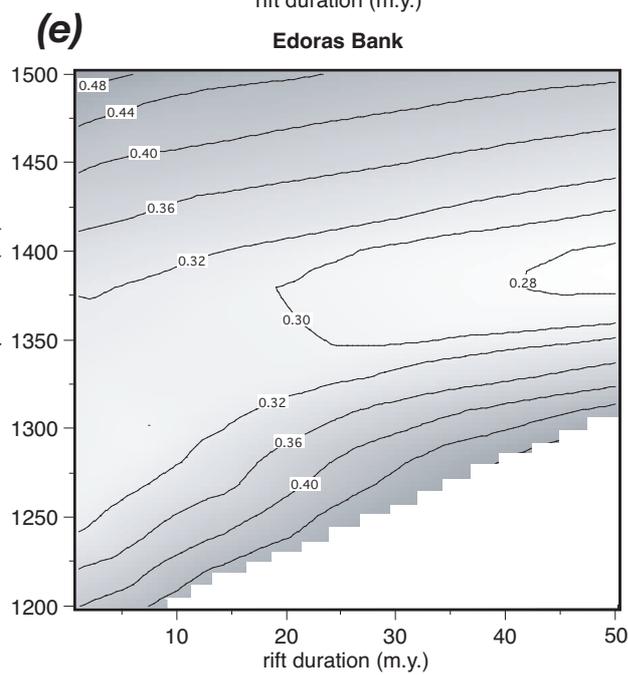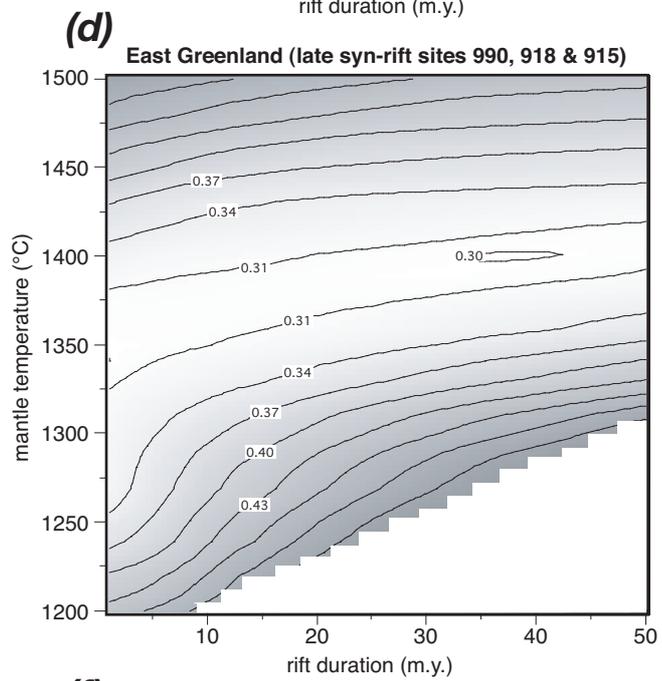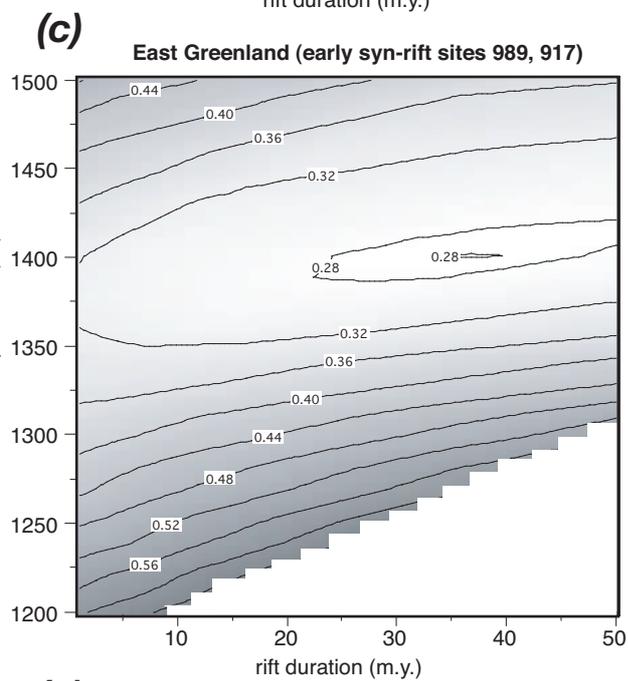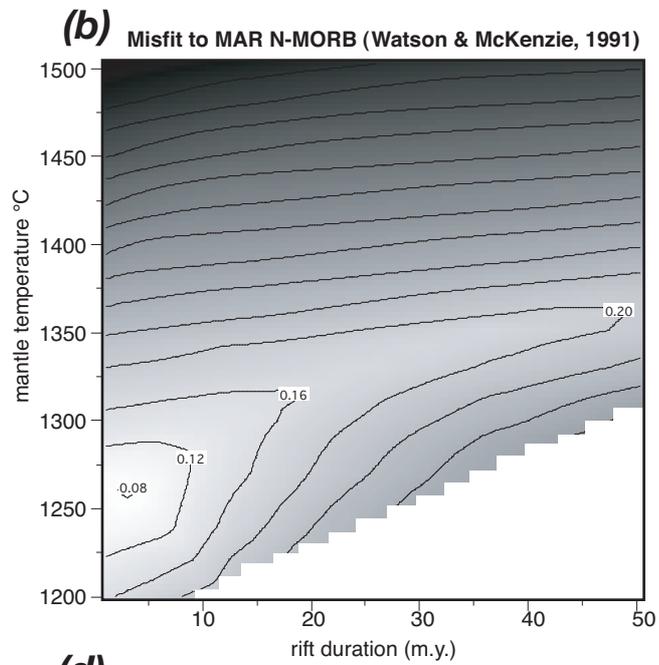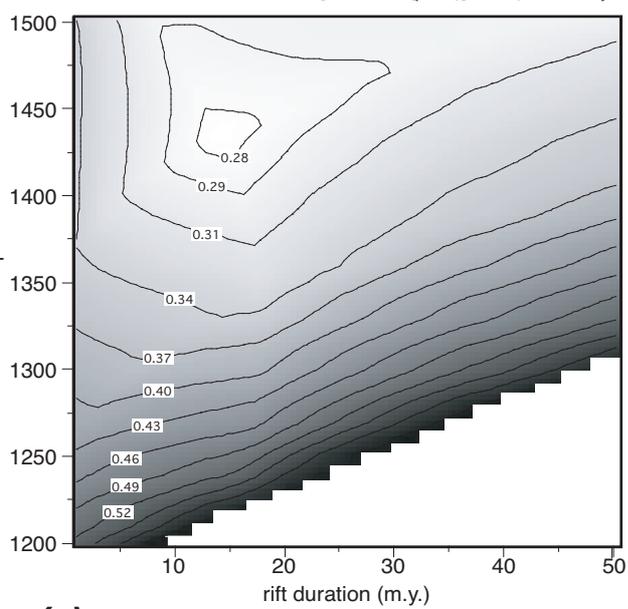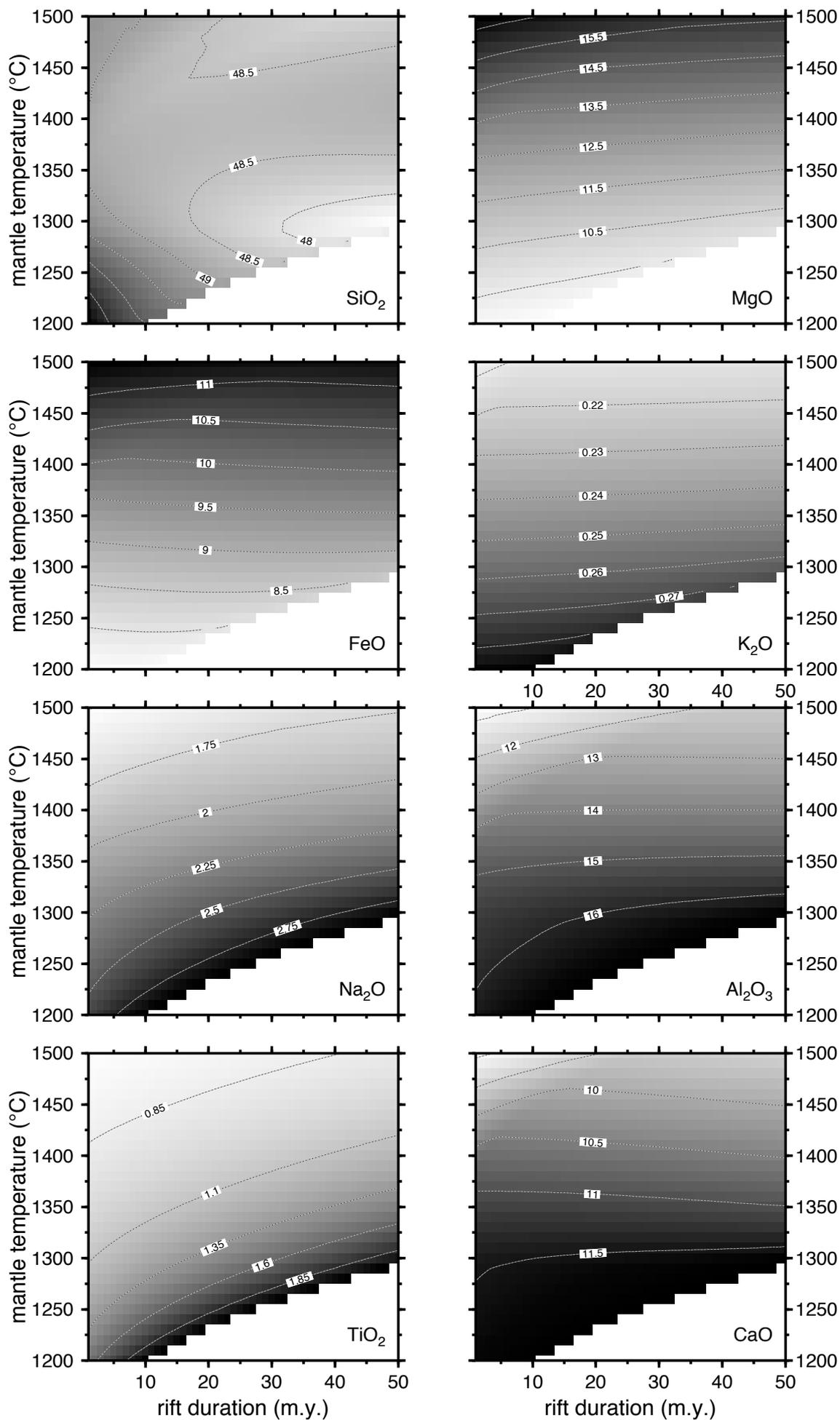
Figure 4
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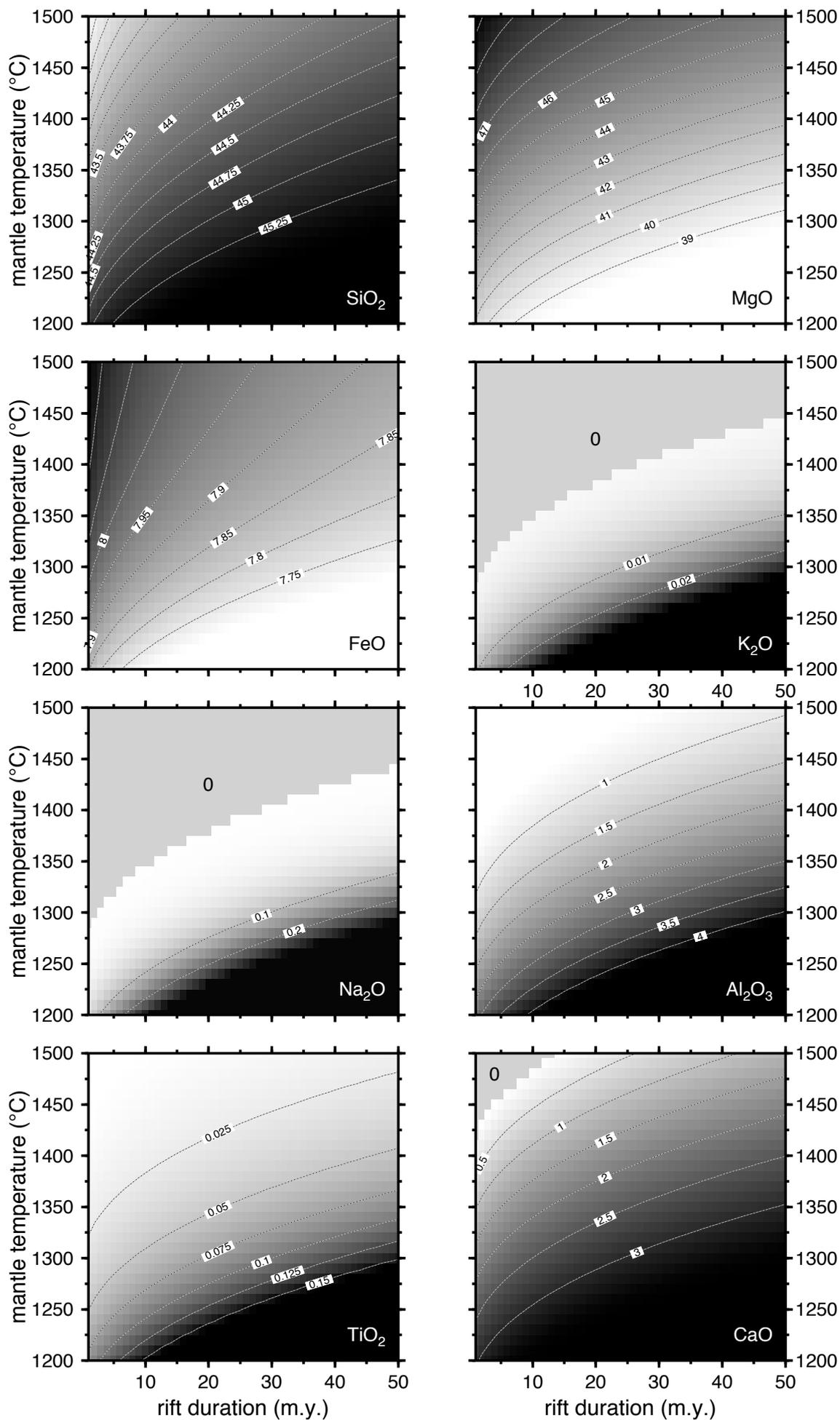
Supplementary material for on-line publication only

Click here to download Supplementary material for on-line publication only: figappendix3.eps

	SiO$_2$	TiO$_2$	Al$_2$O$_3$	FeO	MgO	CaO	Na$_2$O	K$_2$O
Hawaiian pyrolite	45.20	0.71	3.54	8.47	37.50	3.08	0.57	0.13
Tinaquillo lherzolite	44.95	0.08	3.22	7.66	40.03	2.99	0.18	0.02
MORB pyrolite 90	44.74	0.17	4.37	7.55	38.57	3.38	0.40	0.03
Niu (1997)	45.5	0.16	4.2	7.7	38.33	3.4	0.3	0.03

Table 1: Major element source composition from [20] and [21].

Table 2
Click here to download Table: dean_table2.doc

La	Ce	Pr	Nd	Sm	Eu
0.206±30%	0.722±25%	0.143±20%	0.815±20%	0.299±20%	0.115±15%
Gd	Tb	Dy	Ho	Er	Tm
0.419±15%	0.077±10%	0.525±10%	0.12±10%	0.347±10%	0.054±10%
Yb	Lu	Hf			
0.347±10%	0.054±10%	0.22			

Table 2: Concentration (ppm) of rare earth elements for the depleted Earth source of [22].