Porosity evolution of mafic crystal mush during reactive flow

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ABSTRACT

The emergence of the "mush paradigm" has raised several questions for conventional models of magma storage and extraction: how are melts extracted to form eruptible liquid-rich domains? What mechanism controls melt transport in mush-rich systems? Recently, reactive flow has been proposed as a major contributing factor in the formation of high porosity, melt-rich regions. Yet, owing to the absence of accurate geochemical simulations, the influence of reactive flow on the porosity of natural mush systems remains under-constrained. Here, we use a thermodynamically constrained model of melt-mush reaction to simulate the chemical, mineralogical, and physical consequences of reactive flow in a multi-component mush system. Our results demonstrate that reactive flow within troctolitic to gabbroic mushes can drive large changes in mush porosity. For example, primitive magma recharge drives an increase in the system porosity and could trigger melt channelization or mush destabilization, aiding rapid melt transfer through low-porosity mush reservoirs.

MAIN

For most of the last century, models of igneous differentiation and crustal magma transport have been dominated by the hypothesized presence of large, melt-rich magma chambers[1–4]. Within the last few decades, however, there has been a paradigm shift in how magmatic systems are pictured, moving away from the magma chamber hypothesis and towards a model in which magma reservoirs are dominated by crystal-rich mush[5–10].

The importance of mush-dominated magma reservoirs was first recognized at mid-ocean ridges, where geophysical surveys failed to identify the presence of large, melt-rich magma chambers[11,12], and petrological analyses revealed chemical variations that cannot easily be explained by crysatllisation in large, well-mixed magma chambers[13–15]. Petrological and geochemical observations of erupted volcanic products and their cumulate counterparts, from both ocean island and arc volcanic

33   systems, has since confirmed the importance of mush-dominated storage regions worldwide[16–22]. The

34   absence of geophysical evidence for melt-rich magmatic systems (i.e., > 50% melt) in the upper crust

35   beneath active volcanic regions continues to support the mush paradigm[23,24], although the limitations

36   of seismic tomography for imaging thin (<200 m thick) melt-rich bodies has recently been

37   highlighted.[25].

38   The mush paradigm raises fundamental questions about melt transport: if magma plumbing systems

39   are dominated by crystal mush, how is melt transported through them? How does melt accumulate

40   to form eruptible reservoirs? Models for melt extraction or accumulation in mush-dominated systems

41   have traditionally focused on the role of melt buoyancy and crystal compaction, squeezing the more

42   buoyant magma out of the crystalline mush[26–29]. Evidence for crystal compaction is found in the form

43   of plastic deformation of crystals from various magmatic systems worldwide that are shown to have

44   large, long-lived mush reservoirs[17,30]; however, it remains unclear whether magma extraction and the

45   solidification of mush systems can be driven by compaction alone[31,32]. As a result, attention has turned

46   towards the influence of reactive flow, where a melt percolating through a porous crystal framework,

47   possibly aided by mush compaction, reacts with the surrounding crystals ('reactive flow' is here used

48   as a general term to describe both reactive porous flow along grain boundaries and focused flow in

49   melt channels generated by reactive processes)[10,30,33]. Existing numerical models suggest that this

50   process can generate melt-rich domains, channels or layers within crustal mush systems, and that

51   reactive flow may be the dominant mechanism of melt transport and melt extraction in mush-

52   dominated magmatic reservoirs[33,34]. However, current numerical models incorporate highly simplistic

53   empirical chemical parameterizations involving only a single solid composition, which may not

54   accurately recreate the complexity of natural, mineralogically diverse magmatic systems. The

55   influence of reactive flow on the chemical and physical signature in gabbroic, troctolitic, and wehrlitic

56   mush systems has also been investigated through petrological analysis combined with assimilation-

57   fractional crystallization (AFC) calculations[35–37]. However, as AFC calculations are not energy-

58   constrained, they likely violate the conservation of energy principle that governs the behaviour of

59   natural systems (e.g., through the use of thermodynamically infeasible assimilated to crystallised mass

60   ratios – M_a/M_c). As a result, the feasibility of reactive flow as a melt transport mechanism remains to

61   be established, and our understanding of its influence on melt accumulation remains limited.

62   Here, we examine the influence of reactive flow on the porosity of troctolitic to gabbroic mush

63   systems, relevant to mid-ocean ridge magmatism, using thermodynamic simulations implemented

64   through alphaMELTS for MATLAB[38] (rhyolite-MELTS v1.0.2) that capture the full chemical complexity

65   and phase equilibria of natural systems. In addition, by using a Bayesian approach to examine key

66   input parameters (e.g., temperature, initial porosity), we can consider the influence of different

67   variables on the porosity evolution of mafic mushes. This approach also allows us to examine the role

68   of kinetic factors during melt-mush reaction, such as variable mineral dissolution rates[39,40], which

69   might influence the mineralogy of the reacted solid assemblage. Our model results reveal that the

70   influence of reactive flow on the porosity evolution of mafic mush systems beneath active spreading

71   centres is dominantly controlled by the phase proportions in the reacted solid material. In addition,

72   our models highlight key conditions that might drive an increase in the system porosity, triggering

73   mush fluidization or the formation of melt channels, as well as situations where reactive flow causes

74   a decrease in the system porosity, limiting future melt transport and likely driving the extreme

75   chemical enrichment observed in magmatic cumulates[18,41,42]. The thermodynamic approach used here

76   confirms that simple chemical parameterisations used in recent numerical models of reactive flow are

77   insufficient to capture the complex behaviour of natural magmatic systems[33] and highlight that

78   reactive porous flow could be key to melt transport within magmatic mush zones. Overall, our results

79   have substantial implications for the dynamics of magmatic systems in the build-up to volcanic

80   eruptions.

RESULTS

A THERMODYNAMIC APPROACH TO MELT-MUSH REACTION

83   Geochemical and textural signatures of reactive flow have been observed in mafic cumulate rocks

84   from a wide range of tectonic settings worldwide, including mid-ocean ridge[36,41,43], ocean island[18], and

85   arc volcanic systems[22]. However, current chemical and dynamical models of this process lack the

86   ability to accurately determine how reactive flow in a heterogeneous crystal framework influences the

87   porosity of a mush, which hampers our understanding of the implications of reactive flow for the

88   efficiency of melt transport. To address this problem, we simulate the interaction of basaltic melts

89   with olivine + plagioclase ± clinopyroxene mush systems in a series of dissolution-reprecipitation

90   reactions using alphaMELTS for MATLAB (rhyolite-MELTS v1.0.2)[44,45]. The flexibility of our new model

91   enables us to examine the influence of key parameters on mush porosity evolution, including the

92   composition and mineralogy of the initial crystal-framework and the reacted assemblage, as well as

93   the initial porosity and temperature of the system. Our work uses mid-ocean ridge magmatic systems

94   as a template, as the influence of reactive porous flow has been well-documented in these systems,

95   providing a wealth of data to compare to our models[36,37,41,42]. In addition, mid-ocean ridge magmatic

96   systems are typically water-poor and hydrous minerals such as amphibole or biotite are rare, thus

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97   minimizing errors resulting from the absence of accurate thermodynamic data in the MELTS models

98   for these hydrous phases.
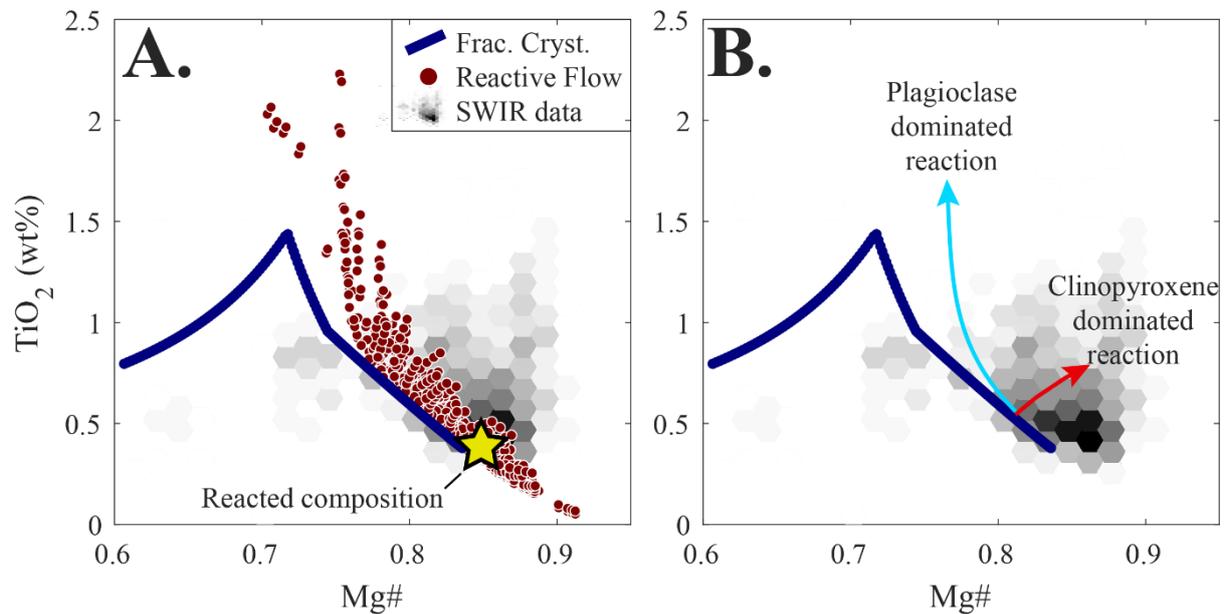
***Figure 1** – Consequences of reactive flow on clinopyroxene compositions. Fractional crystallisation models, using an initial melt composition based on primitive basalts from the Southwest Indian Ridge (see Methods), predict clinopyroxene compositions that bracket the lower TiO₂ contents observed in natural mid-ocean ridge gabbros (data from ref.[36]). Reactive flow models were run over a range of temperatures (1150-1230ºC) with other parameters equivalent to those used in Scenarios 1 and 2 (see Methods and Supplementary Information). **A.** TiO₂ enrichment is observed in many modelled clinopyroxenes, consistent with the natural data. In addition, reactive flow can also cause crystallisation of high Mg# (>0.90) clinopyroxenes – higher than any values observed in crystallisation models or experimental data[56], but in line with the natural data. **B.** Schematic representation of the different paths followed by models dominated by plagioclase and clinopyroxene dissolution.*

99   Initially, to determine whether our models provide an accurate simulation of natural mush processes,

100   we create 200 reactive flow paths over a range of temperatures (1150 – 1230ºC), where a basaltic

101   melt phase reacts with a gabbroic mush zone, and track key chemical parameters over six melt-mush

102   reaction cycles. Results demonstrate that signatures linked to reactive flow in mid-ocean ridge settings

103   are recreated by our models. These include the anomalously high TiO₂ signature and high Mg# (>90)

104   of gabbroic clinopyroxenes, which cannot be generated by fractional crystallisation models alone yet

105   are frequently observed in the rock record[30,36,46] (Fig. 1). Consequently, our energy-constrained

106   reactive flow models, which require far fewer assumptions than traditional Assimilation-Fractional

107   Crystallization calculations[35,36], provide a meaningful simulation of the chemical reactions that occur

108   in natural magmatic mush systems. As such, we can use these models to assess the influence of

109   reactive flow on the porosity evolution of magmatic systems. To do so, we track the Melt Mass Ratio

110   $(M^{Melt}_{Final} / M^{Melt}_{Initial})$ of the local chemical system through the reaction.

111   We consider three scenarios that are appropriate to reactive flow within mid-ocean ridge magma

112   reservoirs. First, we simulate the interaction of a cotectic, three-phase

113   (olivine+plagioclase+clinopyroxene) saturated basaltic melt, which lies along the trend defined by

114   fractional crystallization of a mantle-derived parental magma, with a gabbroic mush (Ol_5:$Plag_{55}$:Cpx_{40}).

115   A second scenario considers higher temperature simulations, where the initial melt phase is only

116   saturated in olivine and plagioclase (i.e., undersaturated with respect to clinopyroxene), but all other

117   parameters are equivalent to those used in scenario 1. Finally, we consider infiltration of 2-or-3 phase

118   saturated basaltic melts into a troctolitic mush system (Ol_{25}:$Plag_{70}$:Cpx_5) over a range of temperatures

119   ($1150 - 1230^{\circ}C$).

120   In each simulation, the initial porosity and temperature of the mush and the melt/solid ratio (defined

121   by setting the mass of reacted, or assimilated, solid material relative to the mass of melt) are randomly

122   selected from a uniform distribution to test their influence on the model results (see Supplementary

123   Information). In addition, by considering the mineralogy of the dissolved solid assemblage separately

124   to the mineralogy of the mush system we can evaluate the role of varying dissolution rates or mineral

125   stability. Additionally, our different scenarios allow the importance of phase saturation in the liquid

126   component to be evaluated.

127   There is abundant evidence in natural magmatic systems that reactive flow proceeds via dissolution-

128   reprecipitation reactions as modelled here[37,41,43]. However, the exact parameters of these reactions,

129   for example, how much dissolution occurs before precipitation is triggered, remain unclear. By

130   considering this problem using a Bayesian approach to sample a broad parameter space for variables

131   such as the mass of reacted material at each step, we can determine the sensitivity of reactive flow to

132   these critical yet under-constrained parameters.

133   REACTIVE CONTROLS ON MUSH POROSITY

134   450 simulations were run for each of the three scenarios. These simulations reveal that during

135   infiltration of three-phase saturated basalts into a gabbroic mush system (i.e., Scenario 1), the modal

136   proportion of clinopyroxene, relative to plagioclase increases in the mush, consistent with reactions

137   proposed based on observations of natural systems[30] and thus further supporting the validity of our

138   models (see Supplementary Information). In fact, many models with low proportions of plagioclase in

139   the reacted solid assemblage ($X_{reaction}^{Plagiocase} < 0.2$) in all three scenarios predict that the dissolved solid

140   material shifts the local chemical systems far enough from the olivine-plagioclase-clinopyroxene

141   cotectic that no plagioclase crystallisation is predicted (Supplementary Information). Nevertheless,

142 there is generally a strong correlation between the proportion of a given mineral in the reacted solid

143 assemblage and the proportion of that phase in the crystallised component (Fig. 2).

Figure 2 – Correlation matrices for each scenario. In all models there is a strong correlation between the Melt Mass Ratio, the proportion of plagioclase in the dissolved/reacted solid assemblage ($X^R_{Plagioclase}$), and the change of specific enthalpy in the melt phase (ΔH: defined as the difference between the specific enthalpy of the melt phase prior to, and after melt-mush reaction). There are, however, subtle differences between the scenarios. For example, in scenario 2, there is a weaker correlation between the proportion of clinopyroxene in the crystallised assemblage ($X^C_{Clinopyroxene}$) and the Melt Mass Ratio than there is in scenario 1. This is likely due to the higher temperature of the scenario 2 models and the clinopyroxene-undersaturated nature of the initial melt phase (and thus lower contribution of clinopyroxene to the crystallising assemblage).

144 We further note that, for the parameters chosen in this study, the Melt Mass Ratio of reaction is

145 typically centred around 0.95 - 1 (Fig. 3). In fact, when the randomly selected dissolved solid

146 assemblage (see Methods) approximates the mineralogy of the mush system, the Melt Mass Ratio is

147 consistently between 0.9 and 1.05. As a result, unless other factors drive the preferential dissolution

148 of one phase relative to the others, which could occur due to differing dissolution rates and/or

149 variations in the melt saturation state, reactive flow often has little to no influence on the porosity of

150 mush systems: dissolution and reprecipitation are close to equal, and reactive flow proceeds by a

151 process akin to zone refining[47,48]. While this result is dependent on other variables, as explored in more

152 detail below, it indicates that reactive flow may operate 'freely', continually working to transport melt

153 through magmatic mush systems and enabling chemical modification of the melt and crystal phases

154 due to the ongoing reactions. However, while the mean Melt Mass Ratio in all scenarios is ~1, the total

155 range of values observed extends from ~0.7 to ~1.3, indicating that under certain conditions, reactive

156 flow can have substantial impacts on mush porosity and thus the efficiency of melt transport.

157 Consideration of all simulations in each scenario reveals that variations in mush temperature, initial

158 porosity, and the solid/melt ratio of the reaction have little influence on the Melt Mass Ratio of

159   reaction (Fig. 2). Instead, our simulations indicate that the mineralogy of the dissolved solid

160   assemblage, here treated independently from the mineralogy of the mush to account for kinetic

161   factors such as variations in mineral dissolution rates[49], represents the dominant control on porosity

162   changes within mush systems. In natural systems, the temperature and melt composition may
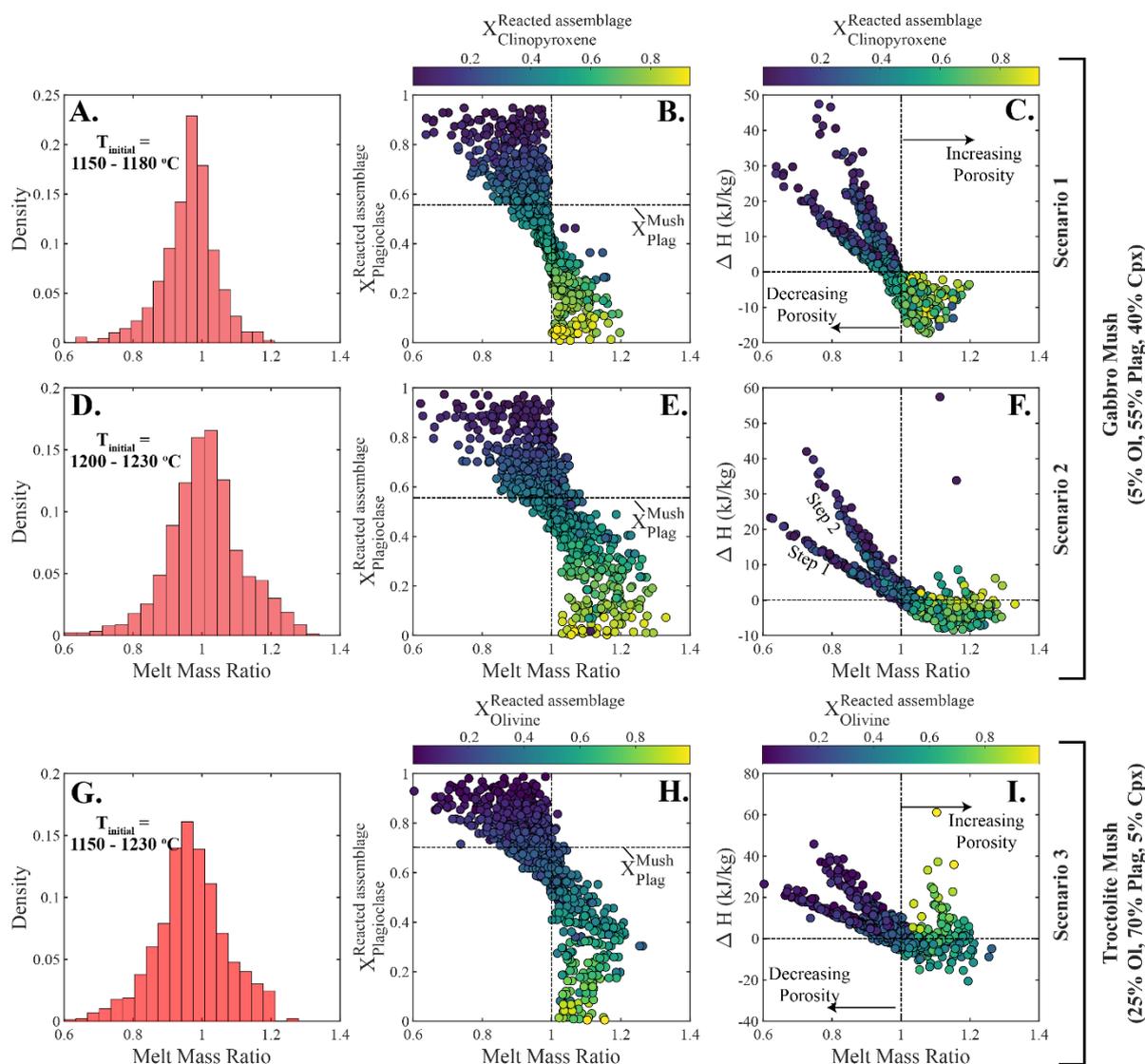
***Figure 3** – Reactive flow controls on the system porosity. The change in the melt mass through a reaction is described as $\frac{Mass_{Final}^{melt}}{Mass_{Initial}^{melt}}$ (Melt Mass Ratio), and is compared to the proportion of plagioclase in the reacted solid assemblage ($X_{Plagioclase}^{Reacted\ assemblage}$) and the change in the specific enthalpy of the melt phase (ΔH). Each simulation (450 per scenario) was allowed to run for 2 steps of melt-mush reaction. **A. - C.** Scenario 1: melt-mush reaction simulations involving an ol+plag+cpx saturated basaltic melt. Results indicate that the Melt Mass Ratio during reactions tend to be clustered around 0.95 - 1 (**A.**). There is a strong negative correlation between the Melt Mass Ratio and the proportion of plagioclase in the assimilated assemblage (**B.**), likely associated with the change in the specific enthalpy of the melt phase (**C.**). **D. – F.** Scenario 2: high-temperature models, otherwise equivalent to scenario 1, where the initial melt phase is saturated in olivine and plagioclase. All results (for a given reacted mineralogy) are shifted to slightly higher Melt Mass Ratios then in Scenario 1. **G. – I.** Scenario 3: the mineralogy of the mush system represents a troctolite, rather than a gabbro. The Melt Mass Ratio increases with decreasing proportions of plagioclase in the reacted assemblage. All models were run at 100 MPa.*

163 influence the mineralogy of the reacted assemblage contributing to the reactions (as discussed

164 below), but these factors are poorly constrained and thus not incorporated into our models.

165 All scenarios reveal a clear correlation between the Melt Mass Ratio and the proportion of plagioclase

166 in the reacted assemblage ($X_{reaction}^{Plagiocase}$). This correlation is driven by the large latent heat component

167 of plagioclase (relative to olivine and clinopyroxene)[50], and the change in the specific enthalpy of the

168 melt phase during plagioclase-dominated reactions (Fig. 3). Specifically, when $X_{reaction}^{Plagiocase} > 0.5 -$

169 0.7, the positive change in the specific enthalpy of the melt phase requires excess crystallisation to

170 balance the enthalpy of the system, and thus drives a decrease in the system porosity. When

171 $X_{reaction}^{Plagiocase} < 0.4 - 0.6$, however, the Melt Mass Ratio is often ~1 or >1, indicating that the porosity

172 is either constant, maintaining the flux of melt through the system, or increasing, enhancing melt

173 transport (Fig. 3).

174 In addition, offsets between scenario 1 and 2 simulations reveal that the phase saturation of the melt

175 component (i.e., the mineral phases that are stable at the liquidus) also plays an important role in the

176 porosity evolution of a mush. Specifically, our models show that interaction of olivine- and plagioclase-

177 saturated, but clinopyroxene-undersaturated melts with a gabbroic mush system can cause an

178 increase in the system porosity as long as $X_{reaction}^{Plagiocase} < 0.6$ (Scenario 2), whereas this transition

179 occurs at ~0.4 for melts saturated in a cotectic phase assemblage of olivine, plagioclase and

180 clinopyroxene (Scenario 1; Fig. 3). The offset between the two scenarios is related to the high latent

181 heat component of plagioclase, which forms a larger proportion of the crystallizing assemblage for

182 models where the initial melt phase is clinopyroxene-undersaturated (Supplementary Information).

183 Consequently, less crystallization (of a plagioclase-rich assemblage) is required to balance the enthalpy

184 of the system. Overall, these results indicate that melt transport through magmatic mush zones can

185 be maintained, or even enhanced by reactive flow in situations where either the melt phase is

186 clinopyroxene-undersaturated or plagioclase does not dominate the dissolved solid assemblage.

187 To investigate this further, we simulate recharge of clinopyroxene-undersaturated melt into a

188 gabbroic mush zone by modifying our model to consider multiple episodes of melt-mush reaction of

189 a single volume of crystal mush. Results confirm that, if there is a sufficient flux of new melt into the

190 mush to maintain clinopyroxene-undersaturation and/or clinopyroxene remains the dominant

191 component in the reacted assemblage, this process will cause a dramatic increase in mush porosity

192 (Fig. 4).

193 However, melt-mush reaction will not always maintain or enhance melt transport: in systems where

194 $X_{reaction}^{Plagiocase} > 0.5 - 0.7$ reactive flow will cause a decrease in the porosity of the system. Critically,

8

195   experimental analysis of mineral dissolution rates indicates that plagioclase dissolution occurs at a

196   faster rate than that of olivine or clinopyroxene, and that the activation enthalpy of dissolution for

197   plagioclase is typically smaller than that of olivine[39,40,49], indicating that this situation may be favoured

198   in many natural systems. Additionally, several petrological studies have shown that key trace element

199   signatures of reactive flow, including the anomalously enriched compositions in clinopyroxene crystal

200   rims[30,36], can be reproduced by models where $X_{reaction}^{Plagiocase} > 0.5$. Therefore, we suggest that

201   plagioclase-dominated reactions – that is, excess plagioclase dissolution relative to the modal

202   proportions of plagioclase in the mush – might be the norm in cooling natural systems (outside of the

203   influence of recharging melts etc.). If so, reactive flow of 3-phase saturated magmas within gabbroic

204   mush systems could be partially responsible for the solidification of magmatic mushes, as well as the

205   generation of enriched rims of low-An plagioclase and and low-Mg# clinopyroxene beneath mid-ocean

206   ridges, reducing the requirement of crystal compaction in generating cumulate rocks from cumulate

207   mush.
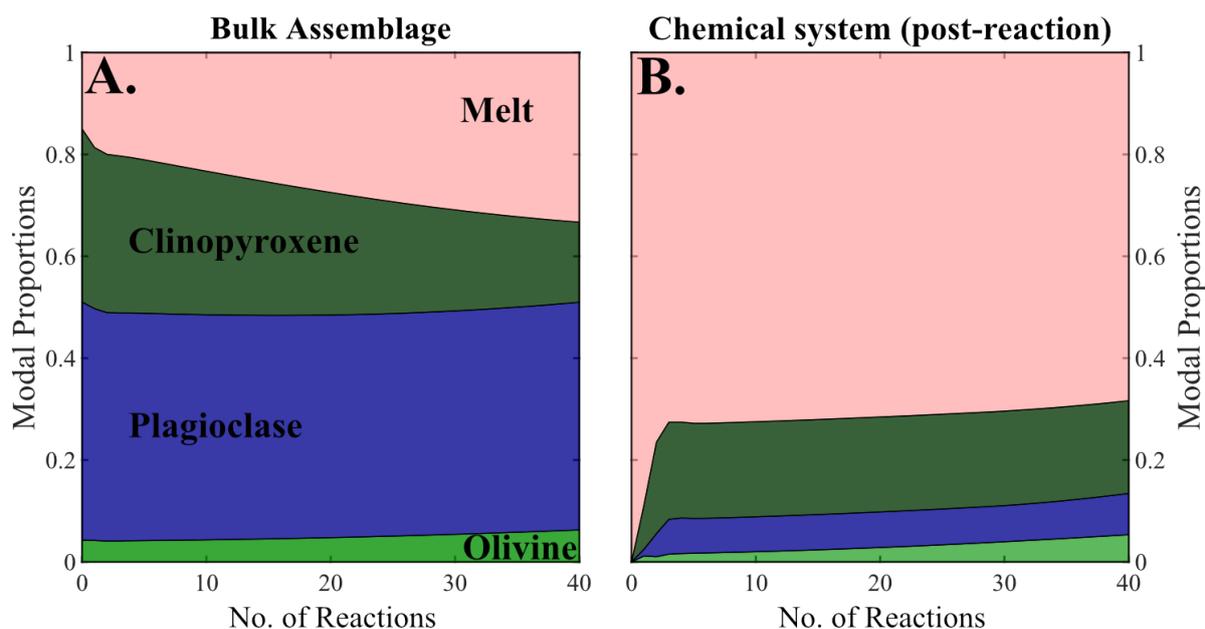
*Figure 4 – Results of melt-mush reaction caused by repeated infiltration of clinopyroxene-undersaturated melts. **A.** Phase proportions in the entire mush system (i.e., melt, reacted solids and unreacted solid assemblage). **B.** Phase proportions excluding solid material that remains chemically isolated from the melt phase (i.e., unreacted solid material). In essence, this panel represents the phase proportions of the reaction at each step of the model. Prior to any model reactions (step 0), the local chemical subsystem is composed entirely of melt. The clinopyroxene-undersaturated nature of the melt phase entering the mush system is assumed to drive clinopyroxene dissolution; as a result, clinopyroxene forms the dominant component within the reacted solid component (0.7) and the melt fraction of the mush increases with each reaction. If this process were to continue, it may be expected that the decrease in crystal-fraction of the mush could eventually lead to partial remobilisation or channelisation within the mush system.*

208 DISCUSSION

209 IMPLICATIONS FOR MELT TRANSPORT IN NATURAL SYSTEMS

210 Compaction of crystal mush, which has long been postulated to be responsible for the formation of

211 eruptible magma bodies and cumulate rocks[33,51], drives porous flow of melts out of the initial mush to

212 accommodate its compaction. Because melts are unlikely to be in equilibrium with the entirety of the

213 solid matrix that they traverse, this porous flow may be reactive. As the Melt Mass Ratio of most

214 reaction is typically centered around 0.95 – 1 (Fig. 3), our findings indicate that porous flow might

215 continue with little-to-no change in porosity and effectively contribute to the formation of melt-rich

216 magma reservoirs and complementary cumulates within crystal-rich systems through compaction.
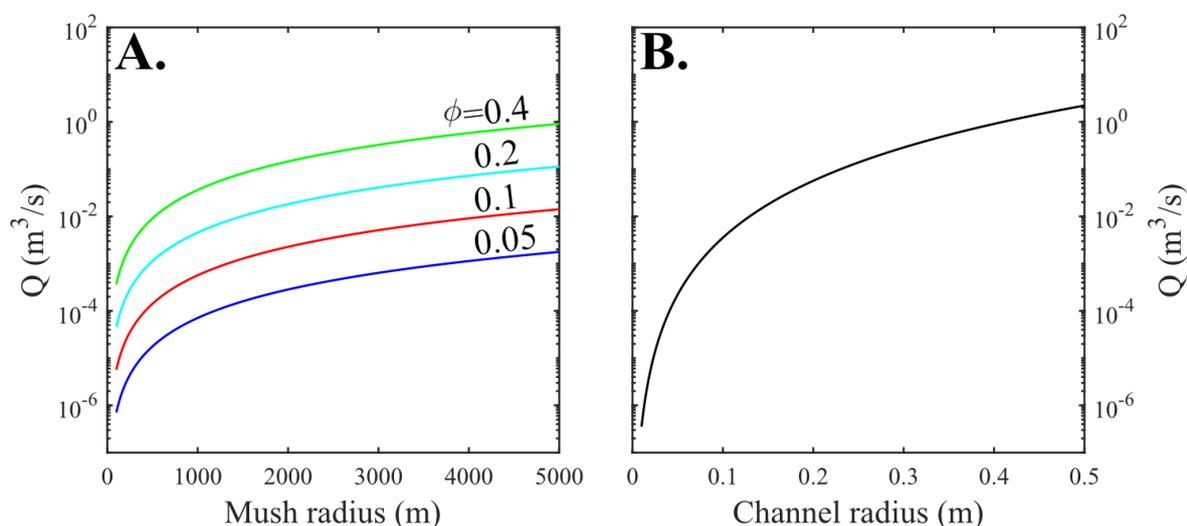
*Figure 5 – Volumetric flux of magma through mush systems by either porous flow (**A.**) or channelized melt flow (**B.**). These calculations demonstrate that channelized melt flow – with channel radii between 10 and 30 cm – can match the volumetric flux of melt through a crystal mush system 1 – 3km in radius with a porosity between 0.1 and 0.4. Volumetric flux calculations are performed using the equations displayed in the Methods section, with phase densities taken from the alphaMELTS for MATLAB fractional crystallisation model at 1180 °C and the mush density calculated using the modal proportions used in scenarios 1 and 2 (ol_5:$plag_{55}$:cpx_{40}). The viscosity of the melt phase is calculated in the Python3 tool Thermobar[57] using the melt viscosity model of Giordano et al.[58] at 1180 °C and the liquid composition from alphaMELTS for MATLAB.*

217 Evidence from natural systems, however, indicates that alternative mechanisms of melt transport are

218 required beyond distributed porous flow. For example, in many fossilized mid-ocean ridge magma

219 reservoirs, erupted lavas and sheeted dykes are in Fe-Mg equilibrium with deep, primitive regions of

220 the mush, but not with the shallower, more evolved sections[10,52,53]. This observation necessitates rapid

221 and efficient transfer of melt through the magmatic mush to avoid magma re-equilibration with the

222 more evolved, shallower gabbros. As such, melt transport by distributed porous flow cannot explain

223 these compositional relationships, indicating that channelized melt flow is required[10]. As dykes are

224 rarely observed crosscutting MOR plutonic sections, this rapid transport must occur at supersolidus

225 conditions, likely in the form of melt channels.

226 Melt channels generated by focused porous flow have been observed on a centimetre- to metre-scale

227 in oceanic core complexes and ophiolites[37,43]. A reactive origin to these channels is supported by their

228 diffuse channel boundaries, mineral compositions indicative of reaction (e.g., clinopyroxene Ti-Cr

229 relationships[30]), and the absence of a shape-preferred orientation in the surrounding crystal mush

230 (indicating limited local deformation driving magmatic flow)[30,37]. Whether such channels also operate

231 on larger scales, like their dunite counterparts in the mantle[54], remains to be determined.

232 Nonetheless, simple back-of-the-envelope calculations confirm that buoyant melt flow within

233 channels is capable, in principle, of transporting significant amounts of magma through magmatic

234 mush systems. For example, the volumetric flux of a buoyant melt within a single 20 cm radius channel

235 is equivalent to that estimated for distributed porous flow through a cylindrical mush system ~3 km

236 in radius with 20% porosity (~0.05 m^3/s; Fig. 5). Our models confirm that reactive flow can drive a

237 dramatic increase in mush porosity and potentially contribute to the formation of these channels.

238 Specifically, increases in mush porosity are likely when clinopyroxene is not on the liquidus and/or

239 olivine or clinopyroxene dissolution occurs at a greater rate than that of plagioclase (Fig. 3). In

240 troctolitic mush systems, olivine dissolution and removal during melt-mush reaction, alongside

241 precipitation of secondary clinopyroxene in channel structures, has been observed[43], and could

242 indicate increasing mush porosity and the formation of melt channels during reactive flow. In gabbroic

243 systems, clinopyroxene-undersaturated melts might drive increased dissolution of clinopyroxene,

244 with both factors acting to increase the system porosity. Therefore, we suggest that an increase in

245 gabbro mush porosity, and potentially the formation of melt channels, is likely during reactive flow at

246 high temperatures, prior to clinopyroxene crystallisation (typically >1180 - 1200°C for MORBs).

247 One scenario that promotes interaction of a high-temperature, clinopyroxene-undersaturated magma

248 with a magmatic mush is replenishment; hence, we suggest that rapid, focused melt extraction may

249 be linked to replenishment episodes. Clinopyroxene dissolution following magma replenishment has

250 recently been documented using high resolution chemical mapping and trace element analysis of large

251 clinopyroxene oikocrysts in the Pacific lower oceanic crust[55]. Our models demonstrate that this

252 process (that is, dissolution of a clinopyroxene-bearing assemblage by a high-temperature,

253 clinopyroxene-undersaturated replenishing melt), combined with the advective and conductive

254 transfer of heat energy into the mush from the new replenishing melt, will drive an increase in the

255 system porosity. Natural variations in the initial mush porosity or mineralogy could then lead to spatial

256 heterogeneity in the rate at which the mush porosity increases, focusing of further melt flow and,

257 ultimately, the formation of a reactive melt channel. Alternatively, dramatic increases in the system

258   porosity surrounding a melt lens formed from a replenishing melt could contribute to mush

259   destabilisation and trigger channelized melt transport (Fig. 6).

260   Once a channel is formed, our models indicate that reactive transport can proceed unimpeded if

261    $X_{reaction}^{Plagiocase}$ remains low (<0.4 - 0.6). However, if $X_{reaction}^{Plagiocase}$ increases, the reactions will drive a

262   decrease in the system porosity limiting future melt transport and possibly generating the extreme

263   chemical enrichment observed in mineral rim zones[30,36,41].

(1) Replenishment. High-T, cpx-undersaturated magma intrudes a gabbroic mush.

(2) Reactive infiltration. Cpx-undersaturated magma ascends into overlying mush and drives an increase in porosity.

(3) Melt focusing. Mush heterogeneity or melt flow instabilities cause focusing of melt flow. Porosity increase is greatest at certain locations.

(4) Channelised flow. Continued melt focusing and mush disaggregation may lead to channelisied melt transport.

Figure 6 – Schematic representation of the processes leading to melt channelization after magma replenishment (1). The replenishing magma infiltrates into the overlying mush and drives an increase in porosity due to the phase stability of the melt and the presence of clinopyroxene in the reacted solid assemblage (2). Heterogeneity in the mush system, or melt flow instabilities could lead to focusing of melt flow into specific region of the mush (3). Melt focusing could ultimately lead to the generation of a melt channel and rapid transfer of magmas within mush reservoirs (4).

264   BROADER IMPLICATIONS

265   Overall, our results demonstrate that reactive flow can have substantial impacts on the porosity of

266   magmatic mush systems, with implications for the efficiency of melt transport and the viscosity of

267   magmatic reservoirs. This work also highlights that more detailed chemical interactions, which

268   account for variations in the mineralogy of the mush and phase saturation of the melt, must be

269 considered in larger, reservoir-scale models of reactive flow and melt transport to fully capture the

270 dynamics of chemically heterogeneous and complex magmatic systems. At present, large-scale

271 models that account for melt and crystal transport in km-scale magmatic systems contain only very

272 simple chemical parameterisations, with all solid material described by a single composition at

273 temperatures above the solidus[33,34]. Our results, however, demonstrate that the heterogeneous,

274 multi-component nature of natural mush systems is critical for accurately modelling the influence of

275 reactive flow and evolution of porosity within magmatic mush systems.

276 ACKNOWLEDGEMENTS

281 DATA AVAILABILITY

282 All code developed in this study is available via the lead-author's GitHub
283 (https://github.com/gleesonm1/MeltMushRxn) and the current version used in this publication has
284 been archived using Zenodo (https://doi.org/10.5281/zenodo.7626521). The GitHub repository is set
285 up with a Results folder containing MATLAB script that will recreate the figures shown in this
286 manuscript, allowing readers to investigate the results of our models in detail.

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426

METHODS

A THERMODYNAMIC SOLUTION TO MELT-MUSH REACTION

We simulate the interaction of basaltic melts with mafic mush systems using the alphaMELTS for MATLAB package and the rhyolite-MELTS v1.0.2 thermodynamic model[44]. Our intention here is not to provide a complete physio-chemical model of melt and crystal transport in magmatic systems, but to construct a thermodynamic framework with which to evaluate the physical consequences of reactive flow.

We use our new code to develop two models of melt-mush reaction. The first considers the evolving nature of a melt phase as it passes through a magmatic system and chemically interacts with multiple regions of mush (Figs. 1 & 3); i.e., the melt composition formed in each reaction proceeds into a new region of uninfluenced mush where melt-mush reaction occurs again. The second model focuses on a single mush horizon and considers the influence of several episodes of melt-mush reaction (due to ongoing melt flow into the mush) on the mineralogy, composition, and porosity of that region (Fig. 4).

To ensure the results of these models are appropriate to natural system, we must consider both the local chemical system within which the reaction occurs (melt + solid reactants) as well as the thermal properties of the wider mush system. This is because, chemical mapping of natural cumulate samples reveals the presence of 'relict' crystal cores that appear to be uninfluenced by melt-mush reaction, indicating that complete chemical equilibrium across the mush system is rarely achieved[30,37]. As a result, we define 2 chemical sub-systems, one characterized by a melt phase and the solid material involved in the reaction (i.e., the dissolved assemblage), and the second characterized by an unreacted solid assemblage. However, as the thermal diffusivity of basaltic melts and mafic mineral phases is several orders of magnitude greater than the chemical diffusivities that control mineral dissolution rates[39,49,59], we assume that the local mush system (including both the unreacted solid assemblage and the active chemical system) remains in thermal equilibrium during melt-mush reaction. Therefore, to determine the new equilibrium assemblage for the mush system following melt-mush reaction we consider both the chemistry of the local subsystems and the energy balance across the mush (i.e., the unreacted solid assemblage may act to buffer any temperature variations caused by melt-mush reaction).

Our new models of reactive flow and melt-mush reaction require several key independent variables to be specified by the user, such as the temperature and pressure of the mush system prior to reactive flow ($T_{initial}$, P). Our model also offers the opportunity to provide a temperature offset between the mush and the percolating melt phase (ΔT), but this is set at 0 for all calculations shown in this study.

459   Other variables include the initial melt mass fraction of the mush system (φ; taken here to broadly

460   represent the mush porosity) and the mass of solid material involved in the reaction relative to the

461   mass of melt in the system (M; i.e., the mass of dissolved solid material), which can also be used to

462   describe the melt/solid ratio of the reaction ($1/M$). These two variables are critical as they can be used

463   to define the relative mass of the three key components within the system: solid material within the

464   mush that remains isolated from reaction ($1 - \varphi * (1 + M)$), the mass of solid material that reacts

465   with the percolating melt ($\varphi * M$), and the mass of melt (φ).

466   The composition of the initial melt phase and the solid phases in the mush system must also be

467   specified (C_{phase}). In addition, it is also necessary to define the mineralogical proportions in the mush

468   (X_{phase}^{Mush}) and the relative mineral proportions in the reacted assemblage (X_{phase}^{React}). Notably, we define

469   these two variables separately, which allows us to investigate the influence of additional factors such

470   as variations in dissolution rates (leading to excess incorporation of a single phase into the reaction).

471   Once these variables have been defined, we can use the rhyolite-MELTS v1.0.2 thermodynamic model,

472   implemented through alphaMELTS for MATLAB, to simulate reactive flow[44,45].

473   To simulate reactive flow, we first perform a simple fractional crystallisation calculation to convert the

474   melt composition provided by the user into a related magmatic composition at the model specified

475   temperature. These fractional crystallisation models initiate at the liquidus of the user-provided melt

476   phase and terminate at the temperature specified for the reactive flow models. At this point, the

477   composition and specific enthalpy of the melt phase, critical for the following calculations, are

478   extracted from the fractional crystallisation models.

479   Next, we use the specified values of M, φ, X_{phase}^{Mush}, X_{phase}^{React}, and the composition of each phase to

480   create 2 chemical systems: the unreacted solid assemblage and the local chemical system of the

481   reaction. In addition, once these chemical subsystems have been defined, we can use the

482   '*calcPhaseProperties*' command in alphaMELTS for MATLAB to calculate the specific enthalpy for each

483   phase in the solid mush; where appropriate, these calculations are performed separately for the solid

484   material that is involved in the chemical reactions and the unreacted solid assemblage. Now that the

485   specific enthalpy of each phase (and each component) within the mush system is known, the total

486   enthalpy of the mush system can be calculated by:

487    $$H^{Mush} = H^{liq} * \varphi + H^{Solid_{unreacted}} * \left(1 - \varphi * (1 + M)\right) + H^{Solid_{reacted}} * M * \varphi$$

488   Where H represents the specific enthalpy of each component (J/kg). Within our models, we assume

489   that there is no flux of heat energy into or out of the system during melt-mush reaction and, as these

490   calculations are performed at a constant pressure, this indicates that the value of H^{Mush} must remain

491   constant during the reaction (where $H^{Mush} = H^{Reaction} * \varphi * (1 + M) + H^{Solid_{unreacted}} * \big(1 - \varphi *$

492     $(1 + M)\big)$). Therefore, any change in the enthalpy of the local chemical system of the reaction

493   ($H^{Reaction}$) must be balanced by changes in the enthalpy (and therefore temperature) of the

494   unreacted solid assemblage ($H^{Solid_{unreacted}}$) and this constraint is critical for determining the new

495   equilibrium assemblage within the mush.

496   In addition to the enthalpy of the mush system, we also need to consider the chemical composition of

497   the local chemical system of the reaction. This is defined as a mixture of the melt phase and the

498   reacted solid assemblage:

499
$$C^{Reaction} = C^{liq} * \frac{\varphi}{\varphi * (1 + M)} + C^{Solid_{reacted}} * \frac{M * \varphi}{\varphi * (1 + M)}$$

500   Once the chemical composition of the reaction is defined, we can proceed to calculate the

501   mineralogical assemblage and phase compositions in the local chemical system following melt-mush

502   reaction. This is achieved using an iterative approach to ensure conservation of energy throughout

503   the reaction (i.e., constant H^{Mush}). First, an initial 'guess' for the post-reaction assemblage is

504   determined at the current temperature of the system and the enthalpy of this new assemblage is

505   recorded ($H_{New}^{Reaction}$). As $H_{New}^{Reaction}$ is unlikely to match the enthalpy of the melt phase and reacted

506   solid assemblage prior to the calculation ($H_{Initial}^{Reaction}$), this initial guess for the post-reaction

507   assemblage will break the constraint that the internal energy of the system (H^{Mush}) must remain

508   constant during the reaction. As a result, we iteratively adjust the temperature of the mush system,

509   and recalculate the post-reaction local equilibrium (and $H_{New}^{Reaction}$) as well as the enthalpy of the

510   unreacted solid assemblage ($H_{New}^{Solid_{unreacted}}$) at each temperature. This process is repeated until

511     $H_{Initial}^{Mush} = H_{New}^{Mush}$, at which point the temperature of the system, alongside the new melt and mineral

512   compositions – as well as their abundances – are recorded. At this point, the process may be repeated,

513   either by taking the new melt composition forward to undergo reaction with a new region of magmatic

514   mush (Model 1; Fig. 1) or by separating the solid fraction of the mush into a reacted and unreacted

515   component, allowing new melt to enter the same region of mush (where the relative fraction of new

516   melt entering the system is defined by a parameter R) and a new reaction to proceed (Model 2; Fig.

517   4).

518   In theory, the model provided in this study is extremely flexible and can easily be expanded to consider

519   other scenarios (e.g., the influence of reactive flow in silicic systems). Simplified MATLAB codes

520   designed to simulate the interaction of basaltic melts with gabbroic mush zones are provided in a

521   GitHub repository and the current version of these models are archived on Zenodo. These models can

522   be easily modified to incorporate different mineral phases, melt compositions, or internal conditions

523   (e.g., temperature and pressure). For the purposes of this study, however, our models are used to

524   focus on the influence of reactive flow in mid-ocean ridge settings. As a result, in all models provided

525   here, we use natural data from the South-West Indian Ridge and the nearby Atlantis Bank oceanic

526   core complex to constrain the composition of the melt and mineral phases in our reactive flow models.

527   Specifically, we use the mean composition of samples KNO0162-9-048-021GL, KNO0162-9-048-004GL

528   and AII0107-6-056-028GL (3 of the most primitive basalt compositions from the SWIR) from ~52 – 55

529    $^{\text{o}}$S on the SWIR as the initial melt composition in our fractional crystallisation and reactive flow models

530   (compositions taken from the compiled database of Gale et al.[60]). Similarly, the composition of the

531   mineral phases used in the reactive flow models are taken from the work of Boulanger et al.[36] on the

532   Atlantis Bank oceanic core complex; in detail, we use the average core composition of minerals

533   measured in olivine-gabbros and troctolites at depths greater than 500 m. Notably, fractional

534   crystallisation models initiated at our proposed starting composition are able to recreate the major

535   compositional trends observed in the SWIR basalts, with reactive flow potentially acting to explain

536   some of the compositional spread around this primary differentiation path (see Supplementary

537   Information).
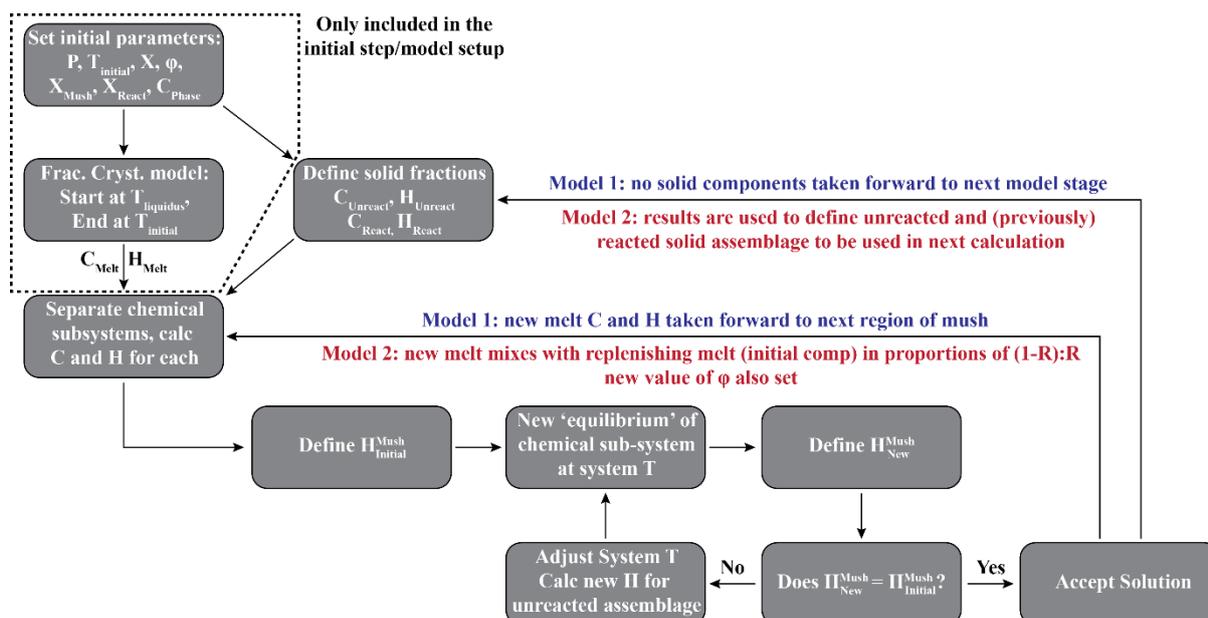
***Methods Figure** – Flow diagram outlining the 2 models developed in this study.*

538   Within the simulations shown in Figs. 1 & 3, we used a Bayesian approach to assess the influence of

539   each parameter on the chemical and physical consequences of reactive porous flow. For all scenarios,

540   the initial values of $T_{initial}$, M, and φ in each simulation were selected from uniform distributions

541   ($0.05 < M < 0.35$; $0.1 < \varphi < 0.35$; T range specified separately for each scenario). The mineralogy

542   of the mush is set to a constant value within each scenario, but the mineralogy of the dissolved solid

543   assemblage is allowed to vary. This is done by creating a random value for each mineral phase ranging

544   from 0 to their modal proportion in the wider mush system and then normalizing the outputted values.

545 For example, in Scenario 1, the code generates random values for olivine, plagioclase, and

546 clinopyroxene between 0 and 0.05, 0.55, and 0.40, respectively. In a single simulation, the code may

547 return hypothetical values of 0.0125, 0.4236, and 0.0890, which when normalised equates to modal

548 proportions in the dissolved solid assemblage of 0.0238, 0.8067, and 0.1695 for olivine, plagioclase,

549 and clinopyroxene, respectively. Which minerals are dissolved by percolating melts in natural systems,

550 and their relative proportions, will likely be influenced by several factors, including the temperature

551 of the system, the relative geochemical enrichment of the percolating melt relative to that which

552 formed in the initial mush (especially as magmas of the lower crust likely retain much of the underlying

553 mantle heterogeneity), and the kinetic barriers to dissolution/reprecipitation. As these factors are

554 difficult to incorporate into our thermodynamic-based models, we believe our Bayesian approach,

555 sampling a wide distribution space, is the most effective means of determining the influence of melt-

556 mush reaction on mush porosity evolution in this study.

557 VOLUMETRIC FLUX CALCULATIONS

558 We provide calculations for the volumetric flux of magma through a mush zone in two different flow

559 regimes. First, we estimate the volumetric flux of magma that can be achieved by buoyancy-driven

560 flow through a high-porosity melt channel. Second, we consider the volumetric flux of magma that

561 results from buoyancy- and compaction-driven flow out of a cumulate mush zone. Comparing the two

562 values provides insights into the role that melt channelization and mush permeability play in the

563 extraction of magma from magmatic mush zones and/or the transfer of magmas within a mush.

564 The volumetric flux of buoyant magma through a high porosity melt channel within a deformable

565 mush zone can be calculated via[61,62]:

566
$$Q_C = \frac{\pi}{8} \times \frac{\Delta\rho g r^4}{\mu_l}$$

567 Where Q_C is the volumetric flux of magma through the channel (m^3/s); $\Delta\rho$ is the density difference

568 between the mush and the ascending melt phase ($\rho_{melt} - \rho_{mush}$; kg/m^3); g represents gravitational

569 acceleration ($-9.81\ m/s^2$); r is the radius of the channel (m); and μ_l is the viscosity of the melt phase

570 ($Pa\ s$). This equation, relating $\Delta\rho$ and r to the volumetric magma flux, is taken from a simplification

571 of plume flow[62], but provides meaningful insights into the volumetric flux of magma that can be

572 achieved in a melt channel if the viscosity of the melt is substantially lower than the medium it is

573 passing through. As the mush has a viscosity several orders of magnitude greater than the melt phase,

574 this condition is met. The calculations are performed using a range of different channel radii, other

575   parameters are taken from alphaMELTS for MATLAB (that is, the density of the different phases) with

576   the exception of the melt viscosity[57,58] (Fig. 5).

577   In contrast, the volumetric flux of magma through a permeable mush reservoir can be calculated via

578   Darcy's Law[63]:

579
$$Q_D = \frac{kA}{\mu_l} \times \frac{\Delta P}{\Delta z}$$

580   Where k represents the permeability of the mush zone (m^2); A is the surface area of the mush zone

581   (m^2); and ΔP is the pressure gradient within the mush over vertical height Δz (Pa/m). The pressure

582   gradient generated by density differences between the melt phase and crystal framework within a

583   mush zone can be written as:

584
$$\Delta z = \frac{\Delta P}{\Delta \rho g}$$

585   Therefore, by rearranging the above equations, the volumetric flux of magma through a permeable

586   mush zone can be written as:

587
$$Q_D = \frac{kA}{\mu_l} \times \Delta \rho g$$

588   Where the permeability of the mush zone is calculated using the method outlined by ref.[34]:

589
$$k = a^2 b \emptyset^\alpha$$

590   Here a represents the typical radius of the grains within the mush zone (m); b is a permeability

591   constant (set to 0.002); α represents a permeability exponent (set to 3); and $\emptyset$ represents the porosity

592   of the mush.

Supplementary Information: Porosity evolution of mafic crystal mush during reactive flow

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Controls on porosity changes

In the main text, we demonstrate that changes in mush porosity during reactive flow are dominantly controlled by the mineralogy of the dissolved solid assemblage. By considering other variables in the melt-mush reaction models we can also examine why these changes are so critical. Correlation matrices for each scenario modelled (shown in the main text and repeated below) confirm that there is a strong correlation between the Melt Mass Ratio and the proportion of plagioclase and clinopyroxene (or olivine for a troctolitic mush) in the dissolved solid assemblage (Fig. 1). However, these calculations also show that the Melt Mass Ratio is correlated with the mineralogy of the crystallised assemblage (unsurprisingly as this is strongly linked to the mineralogy of the dissolved assemblage) and the change in specific enthalpy of the melt phase (ΔH).

Figure 1: Correlation matrices for each scenario. In all models, there is a strong correlation between the Melt Mass Ratio, the proportion of plagioclase in the dissolved solid assemblage ($X_{Plagioclase}^{R}$), and the change of specific enthalpy in the melt phase (ΔH defined as the difference between the specific enthalpy of the melt phase prior to, and after melt-mush reaction). There are, however, subtle differences between the scenarios. For example, in scenario 2, there is a weaker correlation between the proportion of clinopyroxene in the crystallised assemblage ($X_{Clinopyroxene}^{C}$) and the Melt Mass Ratio than there is in scenario 1. This is likely due to the higher temperature of the scenario 2 models and the clinopyroxene-undersaturated nature of the initial melt phase (and thus lower contribution of clinopyroxene to the crystallising assemblage).

In our models, we require that the enthalpy of the entire mush system remains constant (to simulate a thermally equilibrated mush system). As a result, subtle changes in the enthalpy of the melt phase can have significant impacts on the porosity of the mush system. To examine this in more detail, we provide three model runs with plagioclase:clinopyroxene:olivine proportions in the reacted solid assemblage of 75:20:5, 55:40:5, and 25:70:5 (a gabbroic mineralogy is assumed for the wider mush system). Our results indicate that in models dominated by plagioclase dissolution, the difference in enthalpy between the melt phase and dissolved and crystallised solid assemblages increases during melt-mush reaction (Fig. 2). As a result, to maintain the total enthalpy of the mush system, excess crystallisation is required and thus a decrease in the mush porosity. In contrast, when clinopyroxene is dominant, the enthalpy of the melt phase tends towards the enthalpy of the dissolved and crystallised solid assemblages,

meaning that less crystallisation is required to balance the enthalpy of the mush system (and thus an increase in mush porosity).
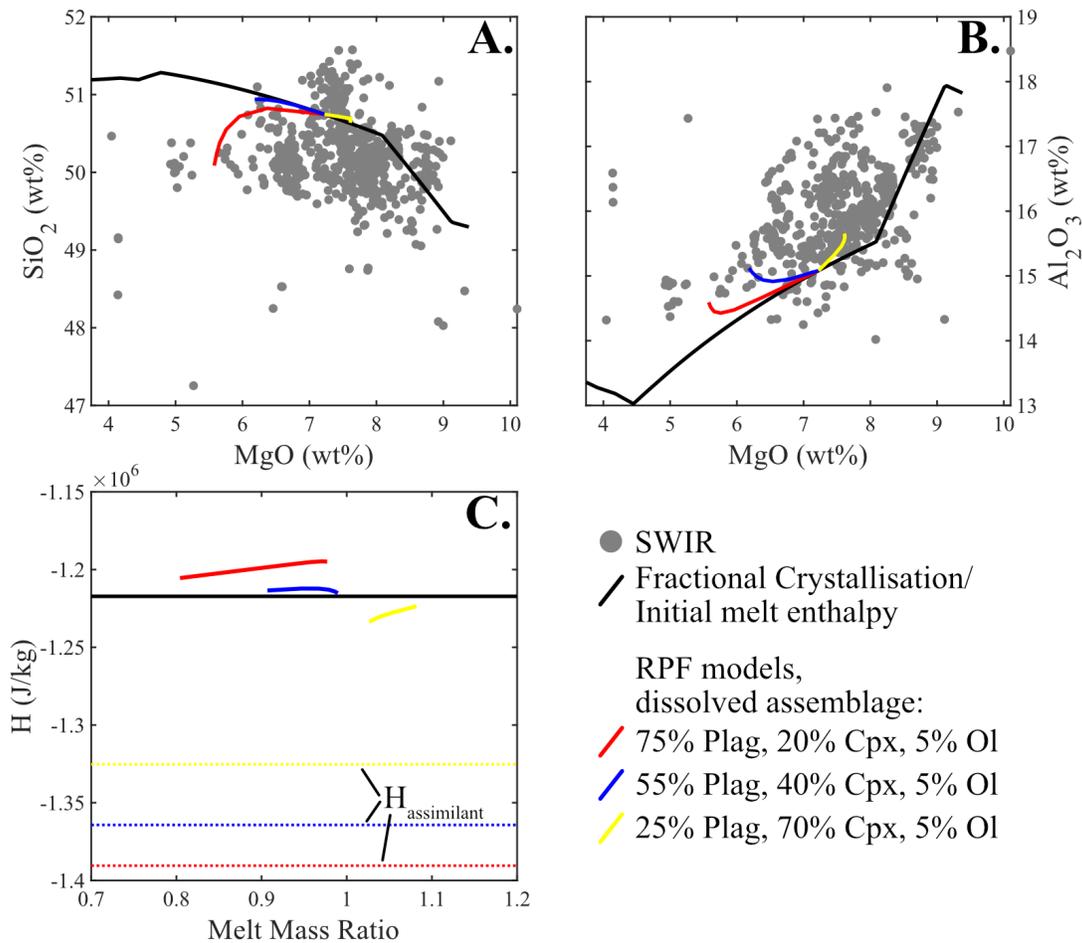
Figure 2: Data shown in **A.** and **B.** represent mid-ocean ridge basalt data of the South-West Indian Ridge from the compilation of Gale et al. (2013). The fractional crystallisation model shown is the same as that used in Fig. 3 of the main text. The reactive flow models initiate at a temperature of 1180 ^oC, with an initial mush porosity of 0.2 and M = 0.3. **A.** and **B.** demonstrate that the melt composition during reactive flow deviates from the path defined by fractional crystallisation. **C.** the difference between the enthalpy of the melt phase and the dissolved solid assemblage increases during melt-mush reaction in models where plagioclase forms a dominant component of the dissolved solid assemblage (>50%). This drives excess crystallisation and a decrease in mush porosity.

Mineralogical changes in the mush

One of the key predictions regarding melt-mush reaction in the literature is that is causes a net change in the mineralogy of the affected mush systems. In fact, several studies have proposed that reactive porous flow in gabbroic mush systems might be characterised by a net increase in the modal proportion of clinopyroxene within the mush system (e.g., equation 1 in Lissenberg et al. 2016).

Using our new models we can evaluate the change in the mineralogy of the mush system as a result of reactive flow and melt-mush reaction. Our results indicate that for the base scenario (i.e., porous flow of a three-phase saturated magma through a gabbroic mush system: Scenario 1), we see a net increase in the proportion of clinopyroxene in the mush system at the expense of plagioclase (Fig. 3). In Scenario 2, where the initial melt phase is clinopyroxene undersaturated, a net increase in the modal proportion of plagioclase in the mush is typically observed, potentially aiding the increase in the porosity of the system found in many of these models due to the high latent heat component of plagioclase (Fig. 3). In addition, our results demonstrate that in Scenarios 1 & 3 there is a net increase in the proportion of mafic minerals (olivine + clinopyroxene) relative to plagioclase in the mush. As a result, when the proportion of plagioclase in the reacted assemblage is low (<0.2), little or no plagioclase is expected in the crystallising assemblage, explaining the peak at 0 in panel **H.** of the histograms below.
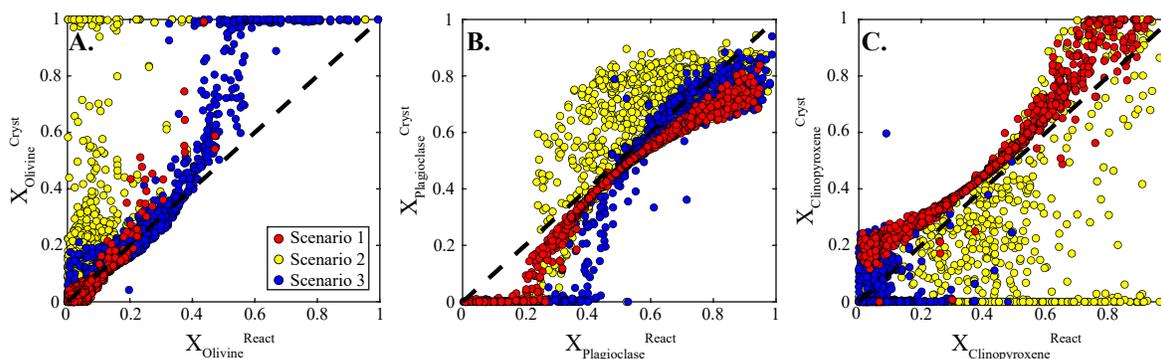
Figure 3: Changes in the mineralogy of the mush, expressed as the proportion of olivine (**A.**), plagioclase (**B.**), and clinopyroxene (**C.**) in the reacted (x-axis) and crystallised (y-axis) assemblage. When data falls above the 1:1 line, this indicates an increase in the proportion of that mineral in the final mush system. For Scenarios 1 & 3, there is a general increase in the proportion of mafic minerals (olivine and clinopyroxene) at the expense of plagioclase in the final mush. In Scenario 2, the clinopyroxene undersaturated nature of the initial melt phases leads to an increase in the proportion of plagioclase in the mush system.

4

Range of input and output values

Below, we show histograms for all three scenarios modelled in this study that demonstrate the range of input variables (in blue) and outputted crystallisation proportions (in red; Figs. 4, 5), 6).

Scenario 1: Gabbro Mush, $T_{initial}$ = 1150 - 1180 ^{o}C
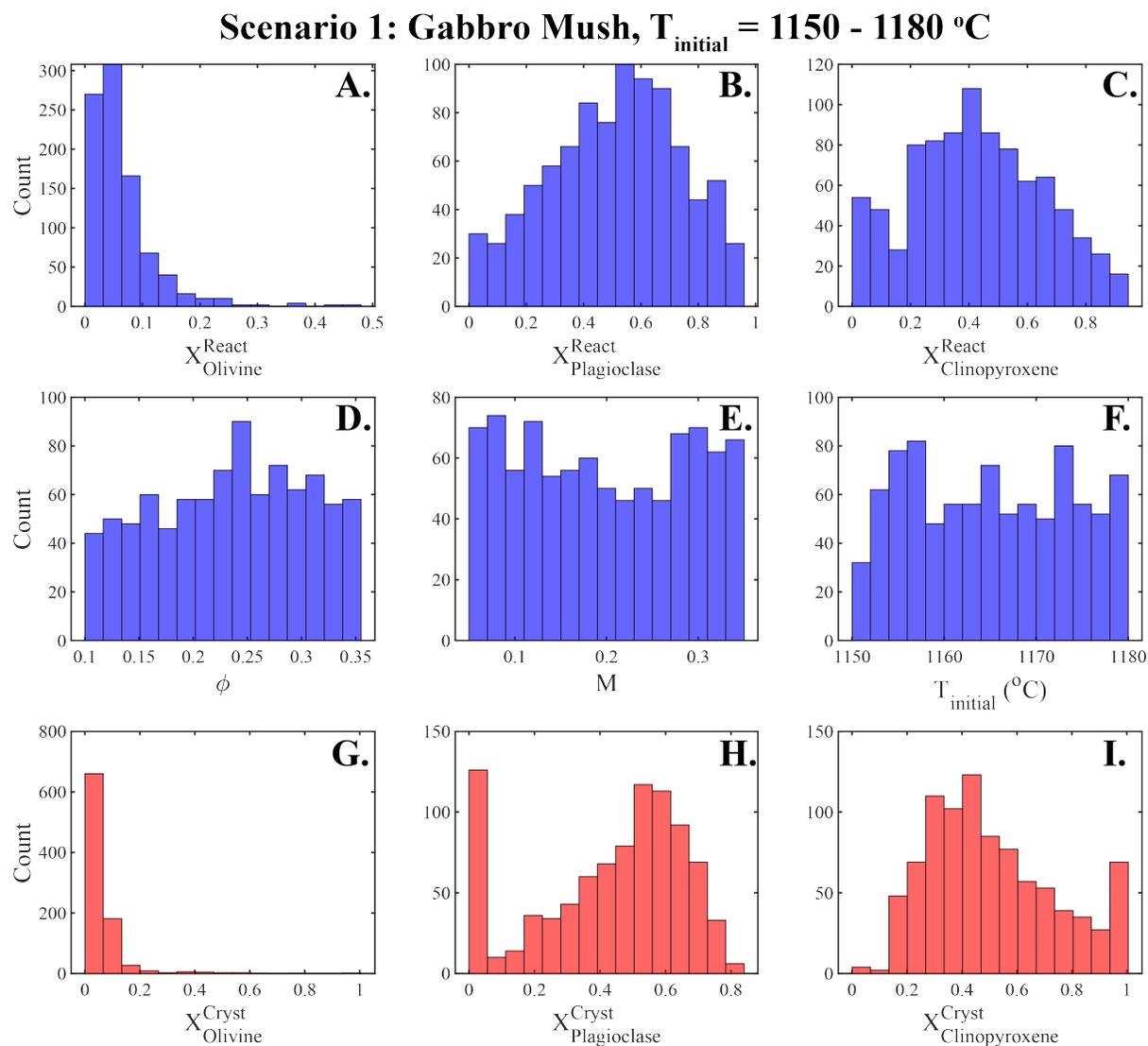
Figure 4: **A. - F.** demonstrate the range of input values used in the 450 simulations (each including 2 steps of melt-mush reaction) displayed in Fig. 3a-c of the main text. **G. - I.** demonstrates the crystallised mineral proportions for these simulations.

Scenario 2: Gabbro Mush, $T_{initial}$ = 1200 - 1230 ^{o}C
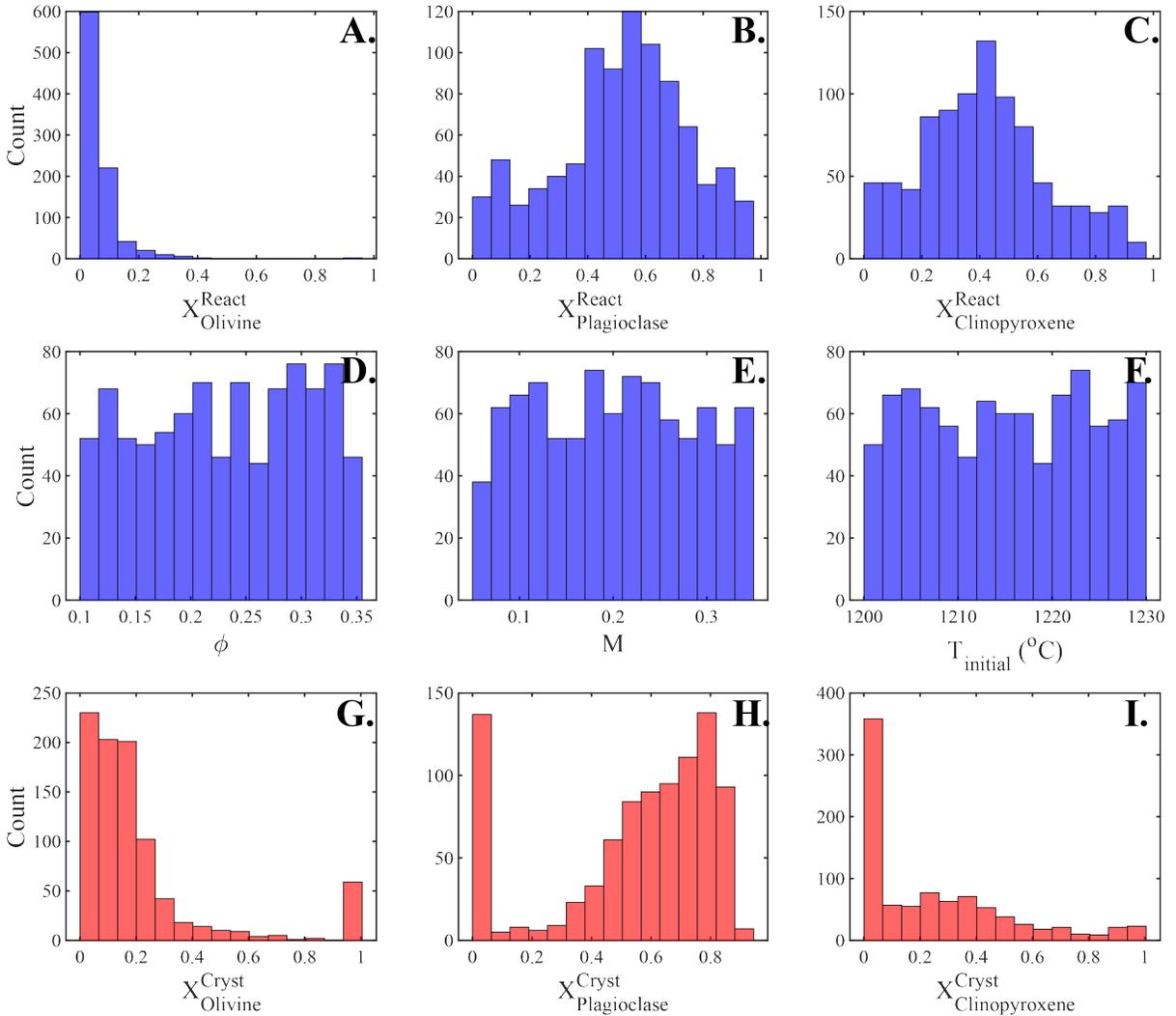
Figure 5: **A. - F.** demonstrate the range of input values used in the 450 simulations (each including 2 steps of melt-mush reaction) displayed in Fig. 3d-f of the main text. **G. - I.** demonstrates the crystallised mineral proportions for these simulations. Notably, compared to Fig. 4, a larger proportion of models show $X_{Clinopyroxene}^{C}$ values <0.2 and $X_{Plagioclase}^{C}$ values >0.6.
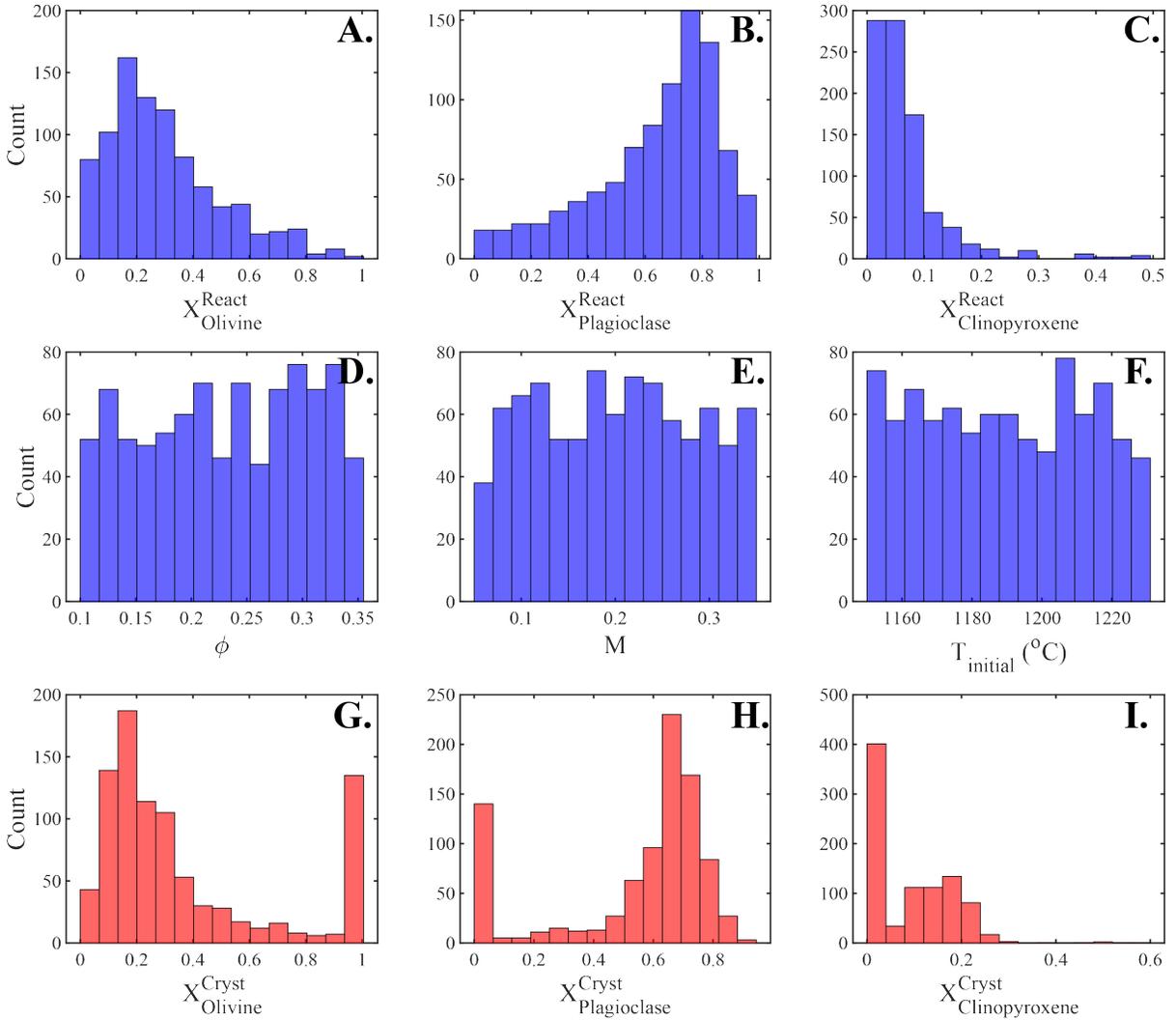
Figure 6: **A. - F.** demonstrate the range of input values used in the 450 simulations (each including 2 steps of melt-mush reaction) displayed in Fig. 3g-i of the main text. **G. - I.** demonstrates the crystallised mineral proportions for these simulations.

Testing different starting compositions

The starting composition used in this study, which is based on the composition of MgO-rich basalts from the South-West Indian Ridge (SWIR) in the compilation of Gale et al. (2013), was chosen for several reasons. First, it represents a reasonable estimate of primary magma compositions beneath the SWIR without having to apply a fractional crystallisation correction. Second, the fractional crystallisation models starting at this proposed primary melt composition provide a good match to the major element data of SWIR basalts, and thus the starting points for our reactive flow/melt-mush reaction models (which are taken from this fractional crystallisation model) are likely appropriate to the SWIR magmas.

Nevertheless, the choice of starting composition could, potentially, have an influence on our model results. To examine this possibility we re-ran all Scenario 1 & 3 models with the melt starting composition set to that of the most primitive Fracture Zone basalt from Dick et al. (2000). Models were not re-run for Scenario 2 as the liquidus temperature of this basalt at the pressure of the models (100 MPa) is only $\sim 1205^oC$. Additionally, only Scenario 3 models with a starting temperature below this liquidus temperature were used in the following figures. Overall, we can see that the choice of starting composition has very little influence on the results of our models (Figs. 7, 8, & 9).
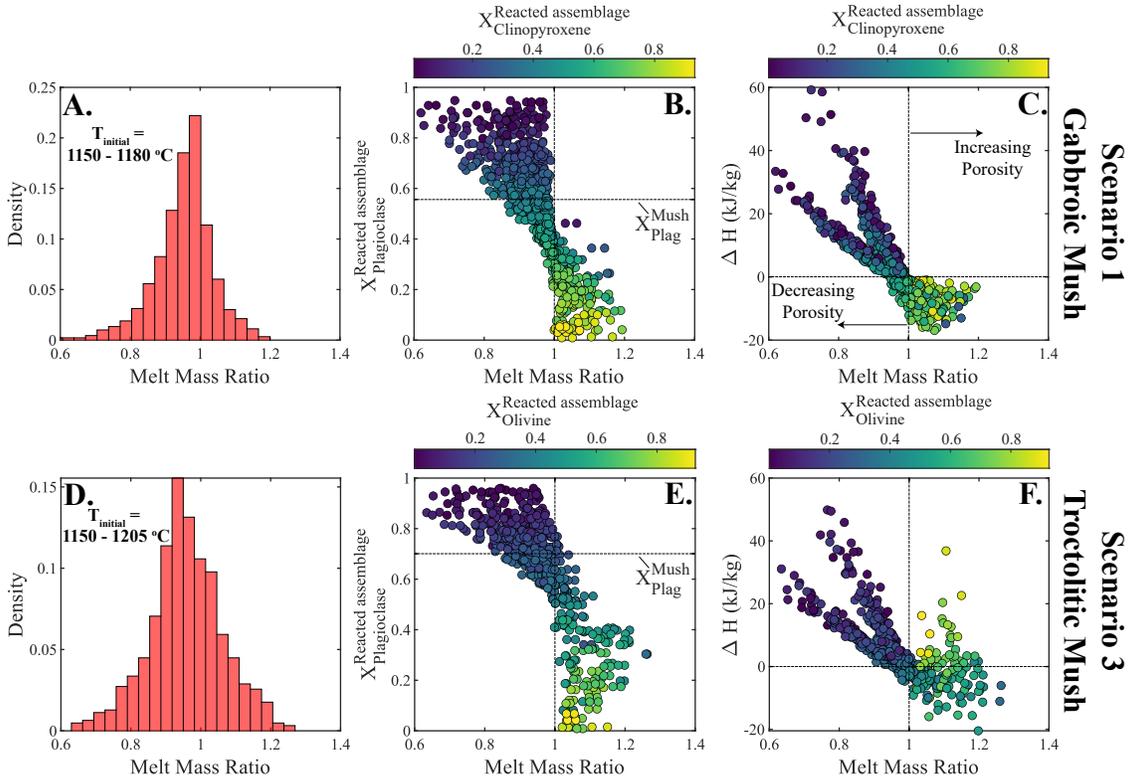
Figure 7: Melt-mush reaction model results for calculations using a starting melt composition from Dick et al. (2000). Scenario 1 & 3 models are shown, and display very similar results to Fig. 3 of the main text.
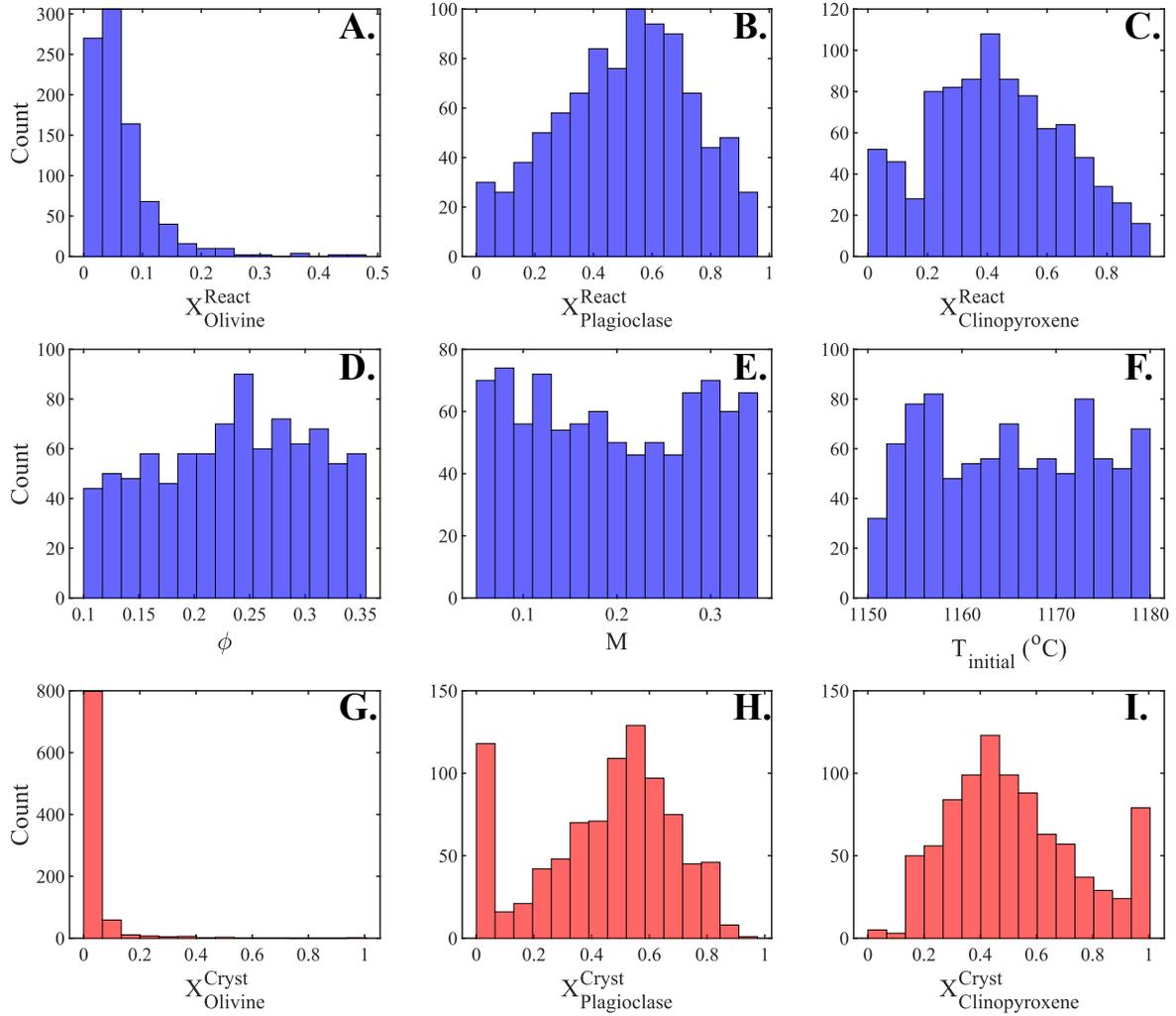
Figure 8: **A. - F.** demonstrate the range of input values used in the 450 simulations (each including 2 steps of melt-mush reaction) displayed in Fig. 7 panels **A.** - **C.**. **G. - I.** demonstrates the crystallised mineral proportions for these simulations.
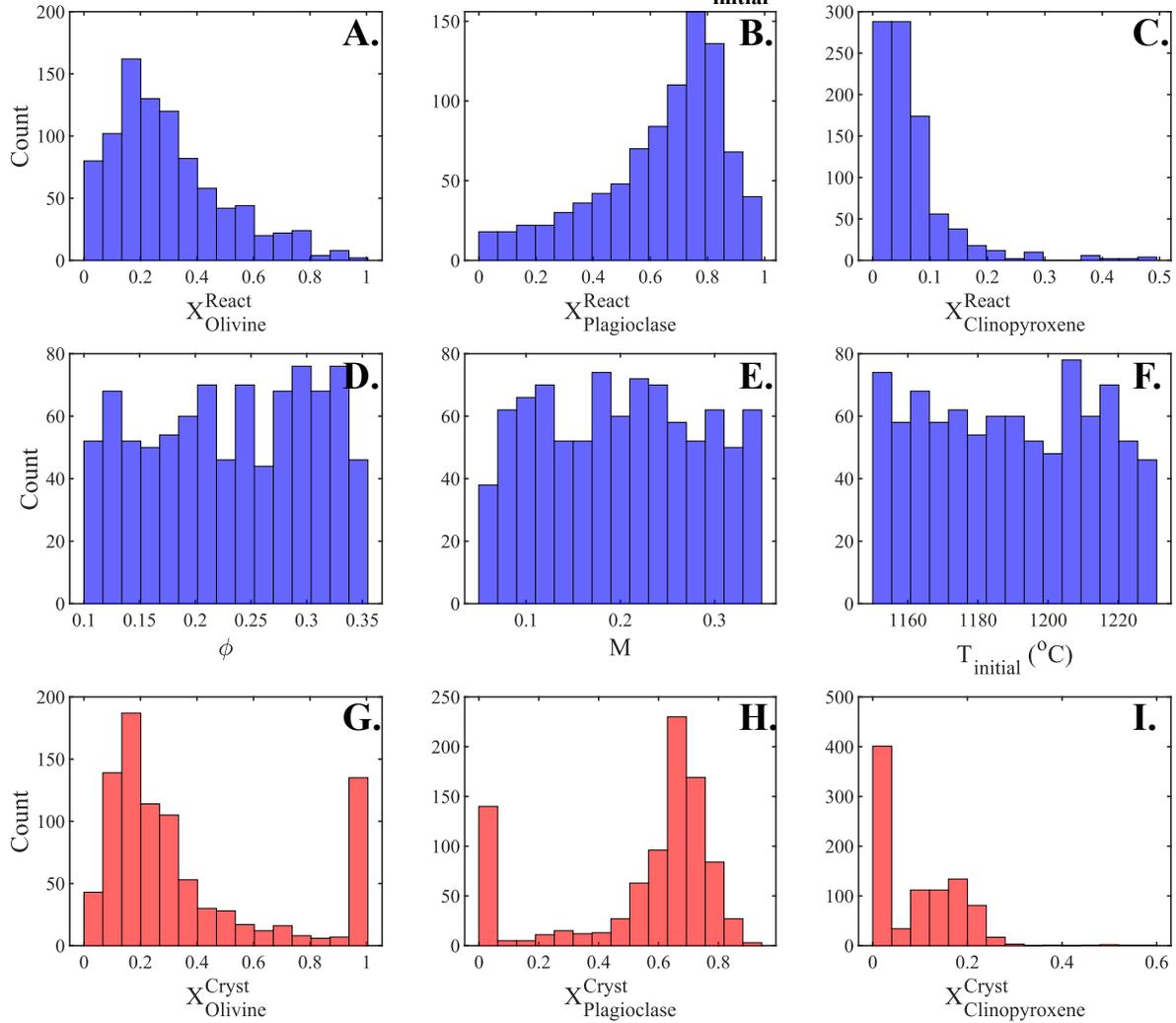
Figure 9: **A. - F.** demonstrate the range of input values used in the 450 simulations (each including 2 steps of melt-mush reaction) displayed in Fig. 7 panels **D.** - **F.**. **G. - I.** demonstrates the crystallised mineral proportions for these simulations.

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