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Key Points:

- Basalt Source Investigation and Characterization (BaSIC) is an open-source Python tool that uses basalt chemistry to constrain apparent mantle source mineralogy and apparent melt fraction
- Tectonic setting and lithospheric thickness exert a primary control on the source mineralogy beneath oceanic islands
- BaSIC detects progressive pyroxene enrichment and decreasing melt fraction during recent eruptions on Iceland's Reykjanes Peninsula

Supporting Information:

Supporting Information may be found in the online version of this article.

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BaSIC: A Bayesian Inversion Model for Basalt Source Investigation and Characterization

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Abstract We present BaSIC: Basalt Source Investigation and Characterization, an open-source Python based Bayesian inversion method to constrain the source mineralogy of basalts. Using bulk rock geochemistry, the model takes the averages and standard deviations of six elemental ratios (Mn/Fe, Zn/Fe, Mn/Zn, Sc/Zn, Co/Fe, and Ga/Sc) for a given basalt suite and inverts for the most likely source mineral assemblage and melt fraction. When applied to a global suite of ocean island basalts, BaSIC reveals a strong tectonic control on apparent source mineralogy. Isolated oceanic islands situated above thick lithosphere yield source assemblages that are systematically poorer in olivine and richer in garnet than islands located near spreading centers or above thinner lithosphere. We attribute this result to preferential sampling of fertile lithologies (rather than cryptic mineralogical variations) in the deeper regions of the melting column. We further test our model using basalts from the recent (2021–2024) eruptions on Iceland's Reykjanes Peninsula. BaSIC detects a progressive pyroxene enrichment in the source and decrease of the bulk melt fraction with successive eruptions, consistent with isotopic studies that show recently erupted lavas are more enriched and isotopically variable than the most recent eruptions (Reykjanes Fires, ~800 years ago) prior to the new cycle. Together, these results demonstrate that BaSIC can resolve both regional-scale variations across ocean island systems and time-evolving source changes within a single volcanic system.

Plain Language Summary Basaltic rocks erupted at ocean islands, such as Hawaii and Iceland, record the chemical “fingerprints” of the rocks from deep inside the Earth that melted to produce them. Certain elements, such as iron, manganese or zinc, are particularly helpful because their ratios in volcanic rock tell us what kinds of minerals were present in the mantle where the melting happened. In this study, we use these elemental ratios to build BaSIC (Basalt Source Investigation and Characterization), a new open-source Python-based numerical melting model. When applied to basalts from 17 ocean islands worldwide, BaSIC reveals that one factor matters more than anything else: the thickness of the overlying rigid lithosphere. Islands residing on thick lithosphere sample a more garnet-rich source, while islands over thinner lithosphere have a source more enriched in olivine. We also tested BaSIC on the recent series of eruptions on Iceland's Reykjanes Peninsula and find that basalts in successive eruptions record an increasing contribution of pyroxene-rich lithologies, while the overall amount of melting decreases. These findings are consistent with recent geochemical studies suggesting that over time these lavas tapped an increasingly enriched reservoir.

1. Introduction

Mantle sourced basalts represent the most abundant and accessible material used to study upper mantle lithologies and their respective chemistries. It has long been recognized that although peridotite is the primary component of the upper mantle, melting of peridotite alone is unlikely to explain the chemical variability of basalts erupted at mid-ocean ridges (MORBs) and ocean islands (OIBs) (e.g., Allègre & Turcotte, 1986; Chase, 1981; Gast et al., 1964; Hofmann, 1997). These findings have prompted the consideration of other lithologies, primarily pyroxenites, as contributors to basalt petrogenesis. Many studies have now shown that including a pyroxenite component in the melting region can more effectively reproduce the observed range of MORB and ocean island basalt (OIB) compositions (e.g., Elkins & Lambart, 2024; Elkins et al., 2019; Hirschmann & Stolper, 1996; Hofmann, 1997; Kogiso, Hirschmann, & Pertermann, 2004; Koornneef et al., 2012; Lambart et al., 2009, 2016, 2019; Mallik et al., 2021; Sobolev et al., 2007; S. Yang et al., 2020).

Here, we use the term pyroxenite to refer to mafic and ultramafic plutonic rocks that lack sufficient olivine (40%) to be classified as peridotites (Le Maitre et al., 2002). Given that pyroxenites can be formed through a large range

of processes, including recycling and delamination of crustal and lithospheric material, in situ fractional crystallization, melt-rock reactions, and solid state reactions (e.g., Downes, 2007; Herzberg, 2011; Lambart et al., 2012; Mallik et al., 2021; Yaxley & Green, 1998), many now agree that pyroxenite comprises at least a small portion of the upper mantle. A number of tracers have been proposed to indicate the presence of pyroxenite in the source region of basalts, including high TiO_2 contents (Prytulak & Elliott, 2007), correlations between major elements and radiogenic isotope ratios (Hauri, 1996; Jackson & Dasgupta, 2008), high FeO combined with low SiO_2 contents (Lambart et al., 2009), low CaO content (Herzberg, 2006), high Fe isotopic ratios (Williams & Bizimis, 2014; Xu et al., 2024), ratios of major elements such as FC3MS (Z.-F. Yang et al., 2016), and log ratios including FCMS and FCKANTMS (Z. F. Yang et al., 2019). However, none of these tracers provide insight into the mineralogical make-up of the mantle source.

First row transition elements (FRTEs) are increasingly popular tracers for constraining mantle lithological heterogeneity with melt geochemistry, as their partitioning behavior throughout melting and fractionation processes allows their source ratios to be preserved more effectively than many other geochemical tracers. Much work now illustrates the potential of FRTEs as geochemical tracers and further characterizes their behavior during melting (e.g., Bowman & Ducea, 2023; Davis et al., 2013; Herzberg, 2011; Howarth & Harris, 2017; Humayun et al., 2004; Lang & Lambart, 2022; Le Roux et al., 2010, 2011, 2015; Mourey et al., 2022; Prytulak & Elliott, 2007; Qin & Humayun, 2008; Sobolev et al., 2005, 2007). First row transition element tracers thought to reflect a pyroxenite contribution to melt include relatively low Mn/Fe , Ni/Co and Mn/Zn , and high Ni/Mn and Zn/Fe ratios (e.g., Bowman & Ducea, 2023; Le Roux et al., 2011; Sobolev et al., 2007). The ratio Ga/Sc has also been suggested as a potential lithological tracer because while Ga and Sc have similar compatibilities in olivine and pyroxenes, they are strongly fractionated by melting in the presence of garnet (Davis et al., 2013). A major limitation of this forward approach is that pyroxenites and peridotites do not constitute two discrete lithologies but instead span a continuous range of mineralogical and chemical compositions (e.g., Herzberg, 2011; Kogiso, Hirschmann, & Pertermann, 2004; Lambart et al., 2013). Consequently, using FRTEs + Ga to distinguish between peridotite- and pyroxenite-derived melts requires assumptions about the mineralogical and compositional characteristics of each endmember that can significantly impact any attempts at quantifying the proportion of the pyroxenite component in the mantle source (Lang & Lambart, 2022).

Recently, Cunningham, Lambart, Guo, et al. (2026) developed a Monte Carlo (MC) statistical model to constrain changes in mineralogy throughout rifting processes on the mid-Norwegian Margin using five elemental ratios (Mn/Fe , Zn/Fe , Mn/Zn , Sc/Zn , and Ga/Sc). Their MC model highlighted a shift in source region mineralogy during rifting, where the peak of magmatism coincided with clinopyroxene enrichment, demonstrating that such models can effectively detect even a discrete shift in source mineralogy.

Here, we further build on the modeling work presented in Cunningham, Lambart, Guo, et al. (2026) with the introduction of BaSIC (Basalt Source Investigation and Characterization). BaSIC is a Bayesian inversion method that expands Cunningham, Lambart, Guo, et al. (2026)'s Monte Carlo statistical melting model by considering Co/Fe in addition to previous ratios (Mn/Fe , Zn/Fe , Mn/Zn , Sc/Zn , Ga/Sc), allowing melt fraction to be a solvable parameter, and by more efficiently exploring the parameter space with the nested sampling Monte Carlo algorithm MLFriends (Buchner, 2014, 2019) within the UltraNest package (Buchner, 2021).

In addition to improving our understanding of mineralogical heterogeneity, outputs from BaSIC can be used to inform other types of geochemical models that have historically relied on assumptions about mantle composition or mineralogy. In the following sections, we outline the use of FRTEs as tracers of mantle heterogeneity, cover the basics of inversion modeling, and demonstrate model application using several suites of OIB data. Documentation to download and run BaSIC is available in the Open Research Statement.

2. Background and Model Construction

2.1. FRTEs as Tracers for Basalt Source Mineralogy

FRTEs (Sc , Ti , V , Cr , Mn , Fe , Co , Ni , Cu , and Zn) and Ga can exhibit contrasted partitioning between individual minerals and melt (D_i), but relatively neutral bulk partitioning behavior (D_0) during melting of mantle lithologies; a characteristic that causes their ratios to be sensitive to the mineralogical assemblage of the source, but minimally impacted by fractionation during partial melting (Cunningham, Lambart, Guo, et al., 2026; Humayun et al., 2004; Le Roux et al., 2010) in comparison to traditional ratios using incompatible elements (e.g., Hertogen &

Gijbels, 1976). Elements of interest here are also not significantly impacted by assimilation of crustal material due to their relatively low abundances in felsic compositions compared to the mantle. Further, it has been shown that changes in mantle temperature similarly impact the partitioning behavior of all FRTEs, meaning that changes in temperature do not strongly impact FRTE ratios (Le Roux et al., 2011). Hence, FRTEs have gained popularity as tracers of mantle heterogeneity, both lithological and mineralogical (e.g., Bowman & Ducea, 2023; Cao et al., 2022; Cunningham, Lambart, Guo, et al., 2026; Davis et al., 2013; Gómez-Ulla et al., 2017; Herzberg, 2011; Howarth & Harris, 2017; Humayun et al., 2004; Le Roux et al., 2010, 2011, 2015; Mallik et al., 2021; Mourey et al., 2022; Prytulak & Elliott, 2007; Qin & Humayun, 2008; Sobolev et al., 2005, 2007; Straub et al., 2008; Xu et al., 2020).

Much work has been done using FRTE concentrations and ratios to differentiate between lavas derived from peridotitic and pyroxenitic source lithologies (e.g., Bowman & Ducea, 2023; Gurenko et al., 2009; Le Roux et al., 2011; Sobolev et al., 2005, 2007). However, the spectrum and overlap in mineralogy space between peridotite and pyroxenite (e.g., Herzberg, 2011; Kogiso, Hirschmann, & Pertermann, 2004; Lambart et al., 2013) requires such studies to rely on assumptions about the nature (composition and mineralogy) of each lithology contributing to melting. Additionally, a single FRTE ratio can be reproduced by a wide range of source mineralogies (Lang & Lambart, 2022). However, the combination of two or more ratios (e.g., Cunningham, Lambart, Guo, et al., 2026; Hole, 2018; Lang & Lambart, 2022; Le Roux et al., 2010; Mourey et al., 2022; Putirka et al., 2011) can be used to reasonably estimate the mineralogical makeup of the mantle source.

2.1.1. Elemental Ratios Used in BaSIC

BaSIC is built upon the MC model from Cunningham, Lambart, Guo, et al. (2026), which used the averaged Mn/Fe, Zn/Fe, Mn/Zn, Sc/Zn, and Ga/Sc ratios and their associated standard deviations from various basalt suites to calculate the mineralogies capable of reproducing those ratios, assuming a constant melt fraction. Each of these ratios were investigated to ensure they were not significantly affected by magmatic processes such as fractionation or crustal assimilation. A more detailed discussion of the choice of the original five ratios in the model is given by Cunningham, Lambart, Guo, et al. (2026). Here, we briefly summarize the main factors influencing this choice.

The authors first identified ratios that are good tracers of mantle source mineralogy, that is, the fractionation behavior of selected ratios must vary as a function of mineralogy. In addition to this, the elemental ratios were selected to fulfill four criteria. (a) Ratios used in the model must include elements that are not dramatically impacted by fractionation processes. In fact, Cunningham, Lambart, Guo, et al. (2026) tested both the effect of olivine and clinopyroxene fractionation (ratio 1:1) and demonstrated that ratios used in BaSIC present relatively low to moderate evolutions down to $F = 0.5$. On the contrary, because Co is less compatible in olivine than Ni (Herzberg et al., 2016), the Ni/Co ratio tends to decrease during fractionation of olivine. In addition, the partitioning of Ni between clinopyroxene and melt is poorly constrained, with values varying from 1.85 (Condamine et al., 2022) to 22 (Pertermann et al., 2004) for experiments performed at 3 GPa. Hence, while Ni/Co has been suggested as a potential tracer for lithological heterogeneity (Le Roux et al., 2011), this ratio was not included in our model. The same reasoning can be applied to other ratios such as Ti/Cr or Ni/Sc (for more details, see Appendix B in Cunningham, Lambart, Guo, et al. (2026)). (b) Source ratios (R_0 ; Equation 1) used in the model must not depend on the potential various lithologies present in the mantle source. For this, elemental ratios were calculated using global suites of natural peridotites and pyroxenites from EarthChem's PetDB database with selected ratios overlapping in the interquartile range (in other words, the middle 50% of two data sets overlap, suggesting a lack of strong statistical difference between their central distributions; see Figure C1 in Cunningham, Lambart, Guo, et al. (2026)). The R_0 and σ_{R_0} values for each ratio used in BaSIC correspond to the average and standard deviation between the respective medians of the peridotite and pyroxenite groups (see details in Appendix C in Cunningham, Lambart, Guo, et al., 2026). (c) Ratios used in the model must not be significantly impacted by secondary chemical alteration processes, such as hydrothermal activity or seawater alteration. BaSIC was developed primarily for use on basalts sampled from oceanic domains, necessitating the consideration of hydrothermal activity and seawater alteration processes. Such processes have been shown to increase the copper content of basalts (Dekov et al., 2013). Hence, ratios involving Cu are not included in the model. (d) Finally, because melt fraction (F) was held constant in the MC model, ratios were chosen for being not significantly affected by the degree of melting up to $F = 0.2$ (see Figure B2 in Cunningham, Lambart, Guo, et al. (2026)).

In BaSIC, we use the five ratios considered in Cunningham, Lambart, Guo, et al. (2026)'s MC model (Mn/Fe, Zn/Fe, Mn/Zn, Sc/Zn, and Ga/Sc) and add Co/Fe. As for the other selected ratios, the contrasting Co-Fe exchange coefficients between minerals and melt make Co/Fe a promising tracer for the mineralogy of the source (Lang & Lambart, 2022; Le Roux et al., 2011). In addition, in Section 2.3, we show that the Co/Fe ratio can help to constrain the degree of melting of the source. However, we also acknowledge that Co/Fe is slightly more sensitive to fractional crystallization than other ratios used in the model (see Figure B1 in Cunningham, Lambart, Guo, et al. (2026)). To further assess the potential effect of fractionation on the outputs of BaSIC, we conducted tests using data from islands with large data sets to assess if there is a systematic shift in outputs when considering only high MgO (5–10 wt.%) versus lower MgO (10–17 wt.%) basalts. If moderate fractionation can affect the results of our model, we should expect to see a systematic shift between the two series across the islands. While results can differ between high and low MgO series for each island (Figure S1 in Supporting Information S1), the lack of systematics shows that the difference cannot be attributed to fractionation. Hence, BaSIC can be applied to samples that have undergone relatively moderate degrees of fractionation, and we recommend using BaSIC only on basalts with more than 5 wt.% MgO, before significant oxide crystallization. We also limit the MgO content to lower than 17 wt.% to prevent the effects of strong olivine accumulation.

2.1.2. Choice of Source Mineral Assemblage

In its current form, source mineralogies considered in BaSIC only contain nominally anhydrous phases (olivine, orthopyroxene, clinopyroxene, garnet, and spinel), despite previous studies arguing for the contribution of hydrous metasomatic phases in the generation of OIBs (e.g., Class & Goldstein, 1997; Niu & O'Hara, 2003; Pilet et al., 2008, 2011). Recently, Ezad et al. (2025) demonstrated that some FRTes can be highly compatible in hydrous phases potentially present in the mantle. In particular, Ni ($D_{\text{Ni}}^{\text{Phl/melt}} = 12\text{--}74.4$) and Cr ($D_{\text{Cr}}^{\text{Phl/melt}} = 5\text{--}161$) are highly compatible in phlogopite, while Sc ($D_{\text{Sc}}^{\text{Amph/melt}} = 0.94\text{--}9.7$), Cr ($D_{\text{Cr}}^{\text{Amph/melt}} = 0.65\text{--}48.6$), Co ($D_{\text{Co}}^{\text{Amph/melt}} = 0.21\text{--}11.5$), and Ni ($D_{\text{Ni}}^{\text{Amph/melt}} = 1.1\text{--}109$) are generally compatible to highly compatible in amphibole. Among these elements, Sc and Co (i.e., those that are included in BaSIC) demonstrate the least variable partitioning behaviors. For these two elements, the average partition coefficients reported by Ezad et al. (2025) for phlogopite and amphibole are $D_{\text{Sc}}^{\text{Phl/lamproite}} = 0.24$, $D_{\text{Sc}}^{\text{Amph/basanite}} = 4.22$, and $D_{\text{Co}}^{\text{Amph/basanite}} = 4.54$. These values suggest that Co is even more compatible in hydrous phases than in olivine (Table S1 in Supporting Information S1). In comparison, while the partitioning behavior of Sc in phlogopite approaches its behavior in olivine, Sc compatibility in amphibole and garnet is similar (Table S1 in Supporting Information S1). Based on these observations, the presence of residual phlogopite in the source may result in an overestimation of the calculated fraction of olivine with BaSIC. Similarly, ignoring the potential presence of amphibole in the mantle source may result in an overestimation of both olivine and garnet fractions. This aspect is discussed in Section 4.2 below. However, without additional experimental constraints, it is challenging at present to quantify the true effect of the presence of hydrous phases in our calculations.

2.2. Modeling Background

Here, we give an overview of Bayesian inversion theory that combines governing petrological equations with sampling tools such as UltraNest (Buchner, 2021). Though not an exhaustive list, we direct the reader to works such as Ullrich et al. (2001) and Tarantola (2005) for more detailed documentation on Bayesian-type inversions and the foundational statistics.

2.2.1. Petrological Foundation

BaSIC relies on elemental ratios measured from basalts (R_B) and their respective standard deviations (σ_{R_B}). Additional model parameters include the partition coefficients for each element (D_i ; Table S1 in Supporting Information S1), and source elemental ratios (R_0). The mathematical basis of the melting model lies in the modal aggregated fractional melting equation, where the elemental ratio of the liquid phase (R_L) for each proposed source mineralogy is calculated as follows (using Mn/Fe as an example):

$$R_{L, \text{Mn/Fe}} = R_{0\text{Mn/Fe}} \times \frac{1 - (1 - F)^{1/D_{0\text{Mn}}}}{1 - (1 - F)^{1/D_{0\text{Fe}}}}, \quad (1)$$

where $R_{L\text{ Mn/Fe}}$ is the Mn/Fe ratio in the resulting liquid phase, $R_{0\text{ Mn/Fe}}$ is the Mn/Fe ratio of the mantle source region, $D_{0\text{ Mn}}$ and $D_{0\text{ Fe}}$ are the bulk partition coefficients for these elements, and F is the melt fraction. In the Monte Carlo approach by Cunningham, Lambart, Guo, et al. (2026), the authors investigated the effect of using variable partition coefficients by letting the D_i values vary at $\pm 1\sigma$ of their published values. They demonstrated that using variable coefficients results in larger dispersion of the outputs but does not significantly affect the relative positions of the outputs between various locations. This can be explained because while D_i are functions of temperature and composition, exchange coefficients (i.e., ratios of partition coefficients for two elements) for the ratios considered here are relatively independent of temperature and composition (Lang & Lambart, 2022; Le Roux et al., 2011; Mallik et al., 2021). Hence, we use constant values for D_i . D_0 values are calculated with the mineral-melt partition coefficients listed in Table S1 in Supporting Information S1 and weighted with respect to the fraction of each mineral (X_{min}) for a given proposed mineralogy.

BaSIC uses modal aggregated fractional melting. Modal fractional, batch, and sequential melting modes were tested during development of the original MC model (Cunningham, Lambart, Guo, et al., 2026), and did not yield significantly different results from fractional melting, consistent with Lang and Lambart's (2022) initial assessment. Because fractional melting is more physically realistic than batch melting, and because continuous melting requires additional assumptions about the melt fraction kept in equilibrium with the source, fractional melting was chosen for use in the standard MC model and is also used in BaSIC. The sensitivity of the model to the modal and non-modal melting was, however, not tested in Cunningham, Lambart, Guo, et al. (2026). We acknowledge that mantle melting is known to be non-modal, with a disproportionate contribution of clinopyroxene to the melt (e.g., Baker & Stolper, 1994; Mallik et al., 2021; Walter, 1998). Non-modal melting requires knowing (or presuming) the melting reaction coefficient for each phase. These coefficients vary with composition (e.g., Pickering-Witter & Johnston, 2000) and pressure (e.g., Krein et al., 2020; Walter, 1998). In addition, non-modal melting models require the implementation of specific mass-conservation equations over given ranges of F to account for the exhaustion of phases due to melting (e.g., Liang, 2024; Liang & Peng, 2010). Because of the nature of very incompatible elements (e.g., REE; HFSE; LILE) in anhydrous mantle sources, traditional melting models (e.g., Hofmann & Class, 2025; Jenner & Jennings, 2025 and references therein) are not very sensitive to the melting mode. In other words, both modal and non-modal melting models produce similar incompatible element concentrations in the melt (Hertogen & Gijbels, 1976). However, because of the relatively neutral compatibility of the elements used in BaSIC, it is important to test BaSIC's sensitivity to melting mode. To do this, we use an approach similar to Lang and Lambart (2022) by comparing the elemental ratios produced by pure peridotite melting and pure pyroxenite melting as a function of the melt fraction obtained using Equation 1 for modal melting, and the following equation for non-modal melting (again using Mn/Fe as an example):

$$R_{L\text{ Mn/Fe}} = R_{0\text{ Mn/Fe}} \times \frac{1 - \left(1 - \frac{P_{\text{Mn}} F}{D_{0\text{ Mn}}}\right)^{\left(\frac{1}{P_{\text{Mn}}}\right)}}{1 - \left(1 - \frac{P_{\text{Fe}} F}{D_{0\text{ Fe}}}\right)^{\left(\frac{1}{P_{\text{Fe}}}\right)}}, \quad (2)$$

where all variables are the same as Equation 1 with the addition of P , the non-modal melting reaction coefficient. To calculate D_0 , we use the same initial mineral proportions as Lang and Lambart (2022): the modal fractions of the peridotite are 0.531 ol, 0.273 opx, 0.177 cpx, 0.019 grt (the subsolidus modal proportion of peridotite KR4003 at 3 GPa; Walter, 1998); the modal proportions of the pyroxenite are 0.82 cpx and 0.18 grt (i.e., the near-solidus modal proportion of the MORB-type eclogite G2 at 3 GPa; Pertermann & Hirschmann, 2003). To calculate P , we use the following melting equations for KR4003 (after Walter, 1998):

$$\text{for } F = 0 - 0.1: \quad 0.083 \text{ ol} + 0.810 \text{ cpx} + 0.298 \text{ grt} = 1 \text{ melt} + 0.191 \text{ opx}, \quad (2a)$$

$$\text{for } F = 0.1 - 0.23: \quad 0.094 \text{ ol} + 1.469 \text{ cpx} = 1 \text{ melt} + 0.563 \text{ opx}, \quad (2b)$$

$$\text{for } F = 0.23 - 0.5: \quad 0.03 \text{ ol} + 0.97 \text{ opx} = 1 \text{ melt}. \quad (2c)$$

For G2, we use:

$$\text{for } F = 0 - 0.5: \quad 0.844 \text{ cpx} + 0.156 \text{ grt} = 1 \text{ melt}. \quad (2d)$$

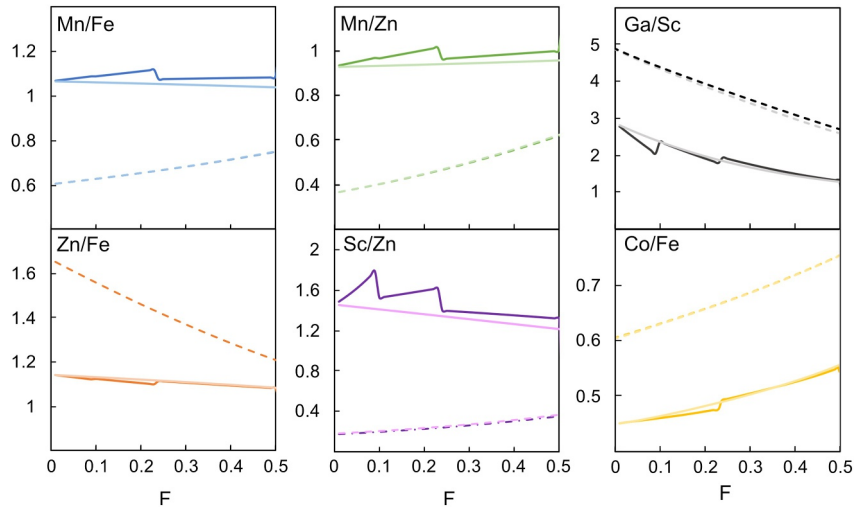


Figure 1. Elemental ratios in the melts produced by peridotite (solid lines) and pyroxenite (dashed lines) as a function of the melt fraction (F) using modal (light color) and non-modal (dark color) fractional melting models.

This melting reaction is derived from experimental results on melting G2 at 3 GPa (Pertermann & Hirschmann, 2003), assuming a linear evolution of the mineral modes for a mineral assemblage with cpx and garnet. P is calculated for each element such that $P = \sum X'_i * D_i$, where X'_i is the melting reaction coefficient of the mineral i in the given forms of Equation 2 (with $X'_i > 0$ and $X'_i < 0$ for reactants and products of the melting reaction, respectively), and D_i is the element's individual mineral/melt partition coefficient (Table S1 in Supporting Information S1).

Aside from the Sc/Zn ratio (and to a lesser extent, Ga/Sc), modal and non-modal melting equations provide very similar results for both lithologies (Figure 1). Sc/Zn for the peridotite-derived melt is more affected by the melting mode, especially at low F when garnet and cpx are contributing to the melt. However, the initial mineral assemblage remains the main control on all ratios, independent of the melting mode. Given the similarities in the results for modal and non-modal melting, we choose to model aggregated modal melting in BaSIC, as it requires fewer assumptions on input parameters.

Finally, the ratio Co/Fe is used in BaSIC both to constrain the mineralogy of the source and to better characterize the melt fraction. For a given melt fraction, $R_{L \text{ Co/Fe}}$ depends on the minerals and their proportions present in the source (Figure 1), providing an additional constraint on the source bulk mineralogy. In addition, unlike for the other ratios, the slope of the Co/Fe ratio with F can be considered as constant for the range of melt fractions considered in BaSIC ($F = 0$ to 0.5; Figure 2a). Hence, we can calibrate the calculated Co/Fe in the basalt as:

$$R_{L \text{ Co/Fe}} = A \times F + R_{L \text{ Co/Fe}}^{F=0} \quad (3)$$

with A (slope) $\approx$ constant and the instantaneous melt at $F = 0$, $R_{L \text{ Co/Fe}}^{F=0}$.

$$R_{L \text{ Co/Fe}}^{F=0} = R_{0 \text{ Co/Fe}} \times \frac{D_{0 \text{ Fe}}}{D_{0 \text{ Co}}} \quad (3a)$$

For 1,000 different assemblages that cover the mineralogical space investigated in BaSIC, we find an average slope $A = 3.35 \times 10^{-4}$. The divergence from the linear trend generally increases when increasing a given melt fraction (F). Applying Equation 3, we solve for the melt fraction required to reproduce $R_{L \text{ Co/Fe}}(F')$, which is calculated using Equation 1. The difference between the given F and F' ($\Delta F = |F' - F|$) tends to slightly increase with given F , but for more than >90% of the data set, $\Delta F \leq 0.05$ (Figure 2b). Therefore, model outputs represent the mineralogies with high likelihood of reproducing all FRTE ratios and melt fractions where $\Delta F \leq 0.05$.

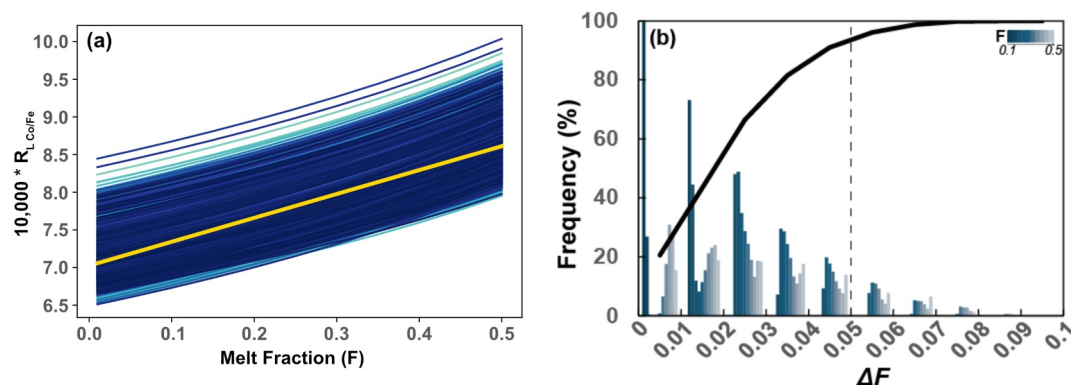


Figure 2. (a) $R_{L\text{ Co/Fe}}$ as a function of F calculated 1,000 mineral assemblages in the mineralogy space covered in BaSIC (blue curves). The yellow line is the linear fit computed with the average intercept ($R_{L\text{ Co/Fe}}$ at $F = 0$) and the average slope (A). (b) Frequency distribution of $\Delta F (=|F' - F|)$. Each series shows the distribution at a given F . The black curve shows the cumulative distribution. 91% of the data are represented for $\Delta F \leq 0.05$.

2.2.2. Bayesian Inversion

Bayesian inverse methods such as the nested sampling employed in BaSIC, explore a range of possible model inputs and evaluate how well the resulting model predictions match observed data, allowing the most likely inputs to be identified. Initial assumptions about the inputs (given by the prior probability distribution) are updated based on their ability to generate models that reproduce the observations, as quantified by the misfit between model results and observations (likelihood function). This process yields a range of plausible solutions that capture both uncertainty in the data and variability in the model inputs (posterior distributions).

Prior probability distributions represent what is known (if anything) about parameter values before considering the observational data. They encode assumptions or constraints on parameter ranges and assign relative probabilities for each parameter, thereby guiding Bayesian sampling. For example, BaSIC calculates mineral assemblages of olivine, clinopyroxene, orthopyroxene, garnet, and spinel. However, not all minerals are allowed to vary freely. Clinopyroxene and orthopyroxene present similar partition coefficients for the elements of interest used in this study (Figure 3). Additionally, except for Mn, the partition coefficients for opx systematically fall between the values for olivine and cpx, making solving for their respective proportions challenging. To limit this issue, we impose an olivine fraction which varies between 0.4 and 1. We believe this is justifiable because, while locally, the presence of lithological heterogeneity might result in olivine-poor to olivine-free assemblages, the bulk mineralogy of the upper mantle (i.e., the region BaSIC aims to constrain) remains largely peridotitic. Additionally, as in Cunningham, Lambart, Guo, et al. (2026)'s MC model, we limit the spinel modal fraction to 0.1 to reflect its accessory nature in ultramafic and mafic rocks. In fact, the typical modal fraction of spinel is 0.5% in abyssal peridotites (Dick & Bullen, 1984), 1%–2.5% in spinel lherzolite xenoliths (e.g., Liang & Elthon, 1990), and 2%–5% in ultramafic massifs (e.g., Gamal El Dien et al., 2019). Spinel is also the most common aluminous phase in ultramafic pyroxenites (Downes, 2007), but its mode is usually limited to <10% (e.g., de Freitas et al., 2026; Laukert et al., 2014). Hence, the prior probability distribution for spinel is represented by a uniform distribution between 0 and 0.1, the prior for olivine is a uniform distribution between 0.4 and 1, and the priors for orthopyroxene, clinopyroxene, and garnet are uniform distributions between 0 and 0.6. We also limit the melt fraction to less than 50%, and thus the prior for F is uniformly distributed between 0 and 0.5.

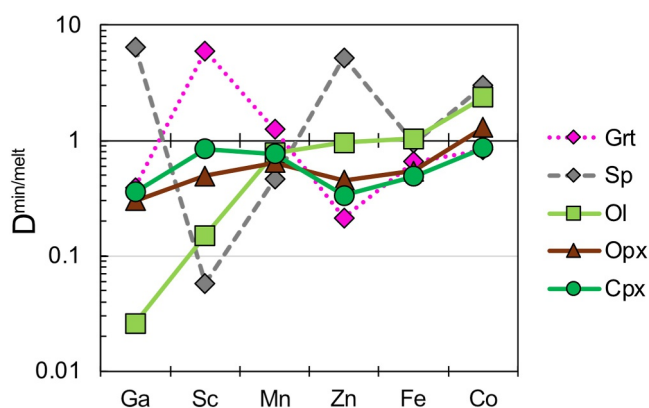


Figure 3. Partition coefficients for the minerals and elements used in BaSIC. Values are from Le Roux et al. (2015). Reported uncertainties in the literature are smaller than the symbols.

To ensure that calculated mineral modes are non-negative and always sum to 1, mineralogies sampled from the prior distribution are drawn from a simplex. A simplex is a geometric shape whose vertices correspond to the end members, and any point within represents a mixture of those end members. Sampling from a simplex is common in modeling practices where there are

dependencies between parameters (i.e., an increase in olivine must be accommodated by a decrease in another mineral). Because only four of the five mineral proportions can vary independently (the sum-to-1 constraint fixes the fifth), mineralogies are sampled from a 4-dimensional simplex, where any point inside the simplex represents a 5-mineral assemblage.

Samples are drawn from the prior distributions and the resulting fit of the model results to the observed data is evaluated using the likelihood function:

$$L = \prod_{i=1}^N \frac{1}{\sqrt{2\pi}\sigma_{R_{B_i}}} \exp\left(-\frac{(R_{B_i} - R_{L_i})^n}{2\sigma_{R_{B_i}}^n}\right), \quad (4)$$

where R_L represents the model outputs, R_B is the measured data (basalt FRTE ratio), σ_{R_B} is the standard deviation of the measured ratio, and $N = 5$ and corresponds to the five ratios used in the model. Following Brown et al. (2020) we employ the least squares misfit to compare the modeled to the observed data (i.e., $n = 2$). To ensure consistency between melt fraction and source mineralogy, any model that does not satisfy Equation 3 with a tolerance on F of $\pm 5\%$, is rejected (i.e., the log likelihood is negative infinity). This is implemented as a model constraint rather than in the prior and thus acts as a likelihood-based filter that enforces consistency between melt fraction and source mineralogy.

2.2.3. Quantifying Posterior Distributions

Nested samplers such as UltraNest (Buchner, 2021) find areas of parameter space with the highest likelihood iteratively, using live points. Each live point represents a unique combination of model parameters (mineral assemblage and melt fraction) drawn from the prior distribution. A BaSIC run begins with 2,000 live points, though the sampler may increase the number of live points at the beginning of a run to better represent the probability space. After this initial increase, the number of live points generally stays constant during the run, but may increase if clusters of probability within the parameter space have been identified during the run. At each iteration, the lowest-likelihood live point is removed (but remembered as a “dead point”) and replaced by a new sample drawn from the prior that is required to have a likelihood greater than that of the discarded point. This imposes a likelihood constraint that progressively restricts sampling to regions of higher likelihood, reducing the fraction of parameter space still consistent with increasingly stringent likelihood thresholds (termed the remaining prior volume, ΔX). The sampler proceeds until the remaining prior volume becomes negligible (i.e., the remaining parameter space is no longer large enough for continued sampling to have a meaningful statistical impact on the results) and the estimated Bayesian evidence (i.e., the integrated model likelihood over parameter space, Z) converges. Each discarded point and each remaining live point at termination is assigned a weight calculated by the product of the likelihood and the prior volume it represents, normalized by the Bayesian evidence ($w_i = L_i \Delta X_i / Z$), allowing posterior distributions of parameters to be constructed by weighted sampling.

An advantage to samplers such as UltraNest (Buchner, 2021) is the decreased reliance on user designated tuning parameters (e.g., method of parameter perturbation and step size), which have the potential to introduce bias into the model. Additionally, nested sampling does not rely on capturing all variability with an initial random scatter shot in parameter space, as was the case in the MC model of Cunningham, Lambart, Guo, et al. (2026). We direct the reader to the recent review from Buchner (2023) for further information on nested sampling (particularly their Figure 5, a nested sampling flow chart).

2.3. Model Outputs

As noted above, UltraNest outputs yield posterior distributions of mineral modes and melt fraction. In addition to these, BaSIC also calculates Aitchison mean values (M_A ; Aitchison, 1986) for the model mineral modes using centered log-ratios of the posterior modes. Weighted geometric Aitchison means are calculated by:

$$M_A = C \exp(\sum_i w_i \ln(x_i)), \quad (5)$$

where C is the closure operator (renormalization to sum to 1), w_i represents the normalized sample weight from UltraNest, and x_i is the proportion of each mineral. Aitchison means appropriately measure the average

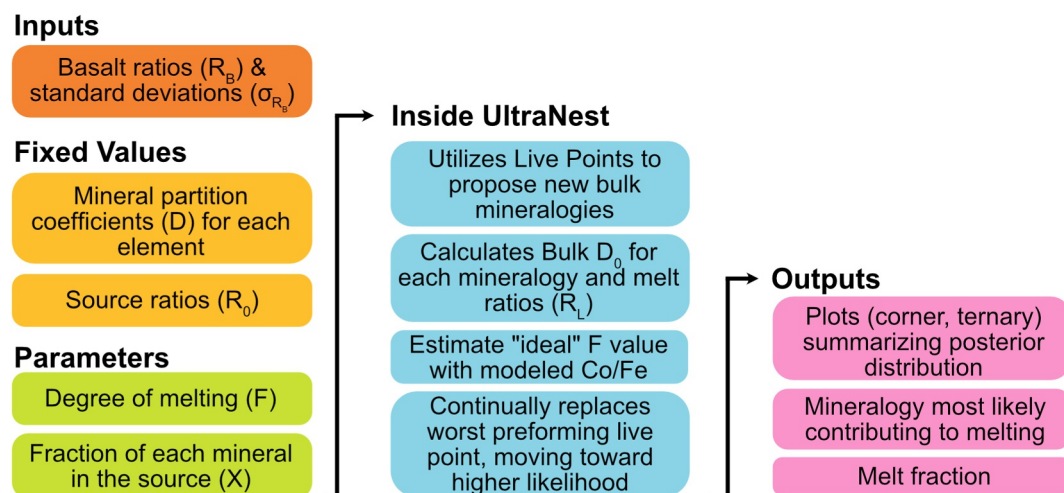


Figure 4. Simplified flow chart depicting BaSIC modeling steps. Inputs, fixed values, and initial parameter values are used in the initial calculations as a starting point for UltraNest. UltraNest calls the function for calculating bulk D_0 and melt first row transition element ratios for each live point. The worst performing live points are continually replaced with points from higher likelihood regions. When updating the live points no longer improves the overall Bayesian evidence, the sampling ends.

mineralogy on the simplex and sum to unity (or 100%). To estimate the uncertainty on M_A , we calculate Aitchison variance (V_A):

$$V_A = \sum w_i D_{Ai}^2, \quad (6)$$

where D_{Ai} is the Aitchison distance, which is a measure of how far an individual posterior sample (i) is from M_A . Low V_A values indicate that the observed elemental ratios can only be reproduced by a narrow range of mineralogies and melt fractions, and therefore, the reported M_A is well constrained. Higher V_A values indicate that the observed ratios can be reproduced by a wider range of mineralogies and melt fractions, meaning that the model outputs are geologically non-unique.

BaSIC outputs include tables containing all posterior mineralogies and their associated melt fractions, a corner plot to visually assess the posterior distribution for each parameter independently, ternary density plots of the entire mineralogy posterior distribution, and ternary plots of the Aitchison (compositional) means. All model outputs presented here were produced with Python 3.9. In addition to MLFriends (Buchner, 2014, 2019) and the UltraNest package (Buchner, 2021), we also used NumPy 1.23.5 (Harris et al., 2020), Pandas 2.2.2 (McKinney, 2010), SciPy 1.10.1 (Virtanen et al., 2020), Matplotlib 3.8.4 (Hunter, 2007), corner.py (Foreman-Mackey, 2016), and MPLtern 1.0.2 (Ikeda, 2024) packages. A flow chart is presented in Figure 4 for a streamlined summary of the model.

3. Application of BaSIC to Oceanic Island Basalts

Due to data availability and relative global coverage, we chose to test BaSIC using a suite of OIB data from the recent OIBID database (Harðardóttir & Jackson, 2025a). Source heterogeneity, both chemical (e.g., Chauvel et al., 1992; Harðardóttir & Jackson, 2025a; Jackson & Dasgupta, 2008; White, 1985; Zindler & Hart, 1986) and lithological (e.g., Hauri, 1996; Hirschmann & Stolper, 1996; Hirschmann et al., 2003; Sobolev et al., 2000, 2005; Soderman et al., 2023) has long been invoked as an explanation for the compositional variability observed in OIBs, along with variations of mantle potential temperature (T_p) (e.g., Herzberg et al., 2007; Mallik et al., 2021), plume source depth (e.g., French & Romanowicz, 2014; Weis et al., 2023), depth of the lithosphere-asthenosphere boundary (LAB; e.g., Dasgupta et al., 2010; Humphreys & Niu, 2009; Jiang et al., 2024; Niu et al., 2011), mantle metasomatism and melt-rock reactions (e.g., Humphreys & Niu, 2009; Mallik & Dasgupta, 2012; Mallik et al., 2021; Niu & O'Hara, 2003; Pilet et al., 2005, 2008, 2011), and the presence of volatiles (e.g., Mallik & Dasgupta, 2013; J. Yang et al., 2025).

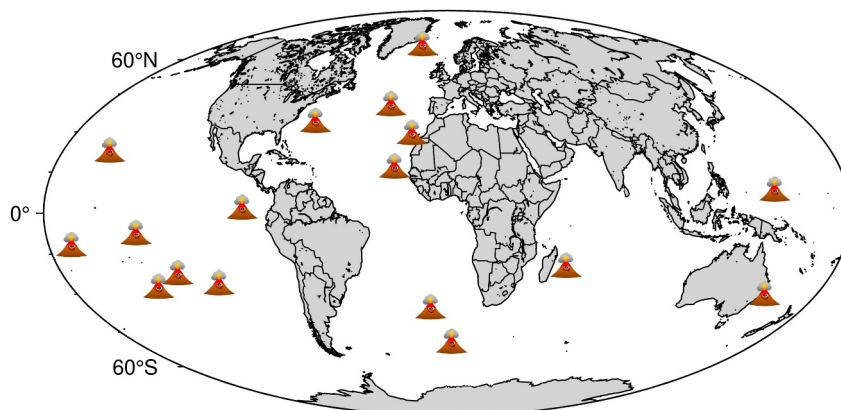


Figure 5. Location of the 17 oceanic islands. Numbers represent islands alphabetically: 1–Azores, 2–Bermuda, 3–Canary, 4–Cape Verde, 5–Caroline, 6–Easter, 7–Galapagos, 8–Hawaii, 9–Iceland, 10–Macdonald, 11–Marquesas, 12–Meteor-Shona, 13–Pitcairn, 14–Reunion, 15–Samoa, 16–Tasmanid-Tasman Central, 17–Tristan-Gough-Walvis Ridge. Map was made using Generic Mapping Tools software in Python (pyGMT; Uieda et al., 2019; Wessel et al., 2019).

The lack of significant felsic crustal influence also makes OIBs an ideal place to test geochemical models that attempt to deduce the nature of the magmatic source. This is not a major concern for BaSIC, since it has been shown that FRTEs are minimally affected by crustal contamination (Cunningham, Lambart, Guo, et al., 2026). However, Cunningham, Lambart, Guo, et al. (2026) showed that their MC model is sensitive to re-entrainment of mafic melts that had previously stalled in the thinning lithosphere. Hence, the intra-plate, and therefore relatively tectonically quiescent, setting of many OIBs allows for testing of the model without concern for the influence of lithospheric thinning.

Data were filtered to include only basalts with more than 5 and less than 17 wt.% MgO and for which all elements used in BaSIC (Mn, Fe, Zn, Sc, Ga, Co) are provided. We also removed significant outliers from each suite by calculating a 3 standard deviation envelope for the basalts from each island. Analyses with three oxides or elements (or more) outside of the 3 standard deviation envelopes were discarded before calculating the averages and standard deviations used as model inputs. Finally, we only selected sample suites that contain 10 or more basalt analyses after filtering. This resulted in the following OIB suites: Azores ($n = 74$), Bermuda ($n = 17$), Canary ($n = 57$), Cape Verde ($n = 38$), Caroline ($n = 22$), Easter ($n = 35$), Galapagos ($n = 66$), Hawaii, ($n = 100$), Iceland ($n = 40$), Macdonald ($n = 15$), Marquesas ($n = 12$), Meteor-Shona ($n = 11$), Pitcairn ($n = 20$), Reunion ($n = 79$), Tristan-Gough-Walvis Ridge ($n = 50$), and Tasmanid-Tasman Central ($n = 11$) (Figure 5).

Although we use OIBID data for general Hawaiian basalts, information about the stage of volcanism is not always included. With the abundance of basalt geochemical data in Hawaii, it is conventional to use only shield stage lavas when modeling the source (e.g., Jackson et al., 2012). To model shield stage only lavas, we supplement OIBID with a separate data set compiled from the GEOROC database (access date: 14 April 2025). For Iceland, we also supplement OIBID with additional data from the most recent Reykjanes Peninsula eruptions (Bindeman et al., 2022; Troll et al., 2024).

While samples from Samoa are included in the OIBID data set, none had all the required trace element data to be used in BaSIC. The EartChem database (PetDB; access date: 28 January 2025; $n = 3$) includes only three analyses for which all five elements are reported. While this does not meet the threshold of 10 analyses mentioned above, the nature of the mantle source of Samoa has been the subject of many studies (e.g., Adams, Jackson, et al., 2021; Jackson et al., 2007, 2010; Wang et al., 2021; Wright & White, 1987), and the low $^3\text{He}/^4\text{He}$, high $^{87}\text{Sr}/^{86}\text{Sr}$ component of Samoa is considered an end-member mantle composition termed EMII (Adams, Spera, & Jackson, 2021; Workman et al., 2004). Hence, we decided to test our model on this small OIB suite. Finally, we include calculations performed on the average N-MORB composition (Gale et al., 2013) as a reference point.

Selected modeled source mineralogies are summarized in density plots (Figure 6 and Figure S2 in Supporting Information S1). An example of the calculated mineralogy and melt fraction distributions is presented for Iceland in Figure 7. Similar corner plots for each individual island are available in Figures S3–S24 in Supporting

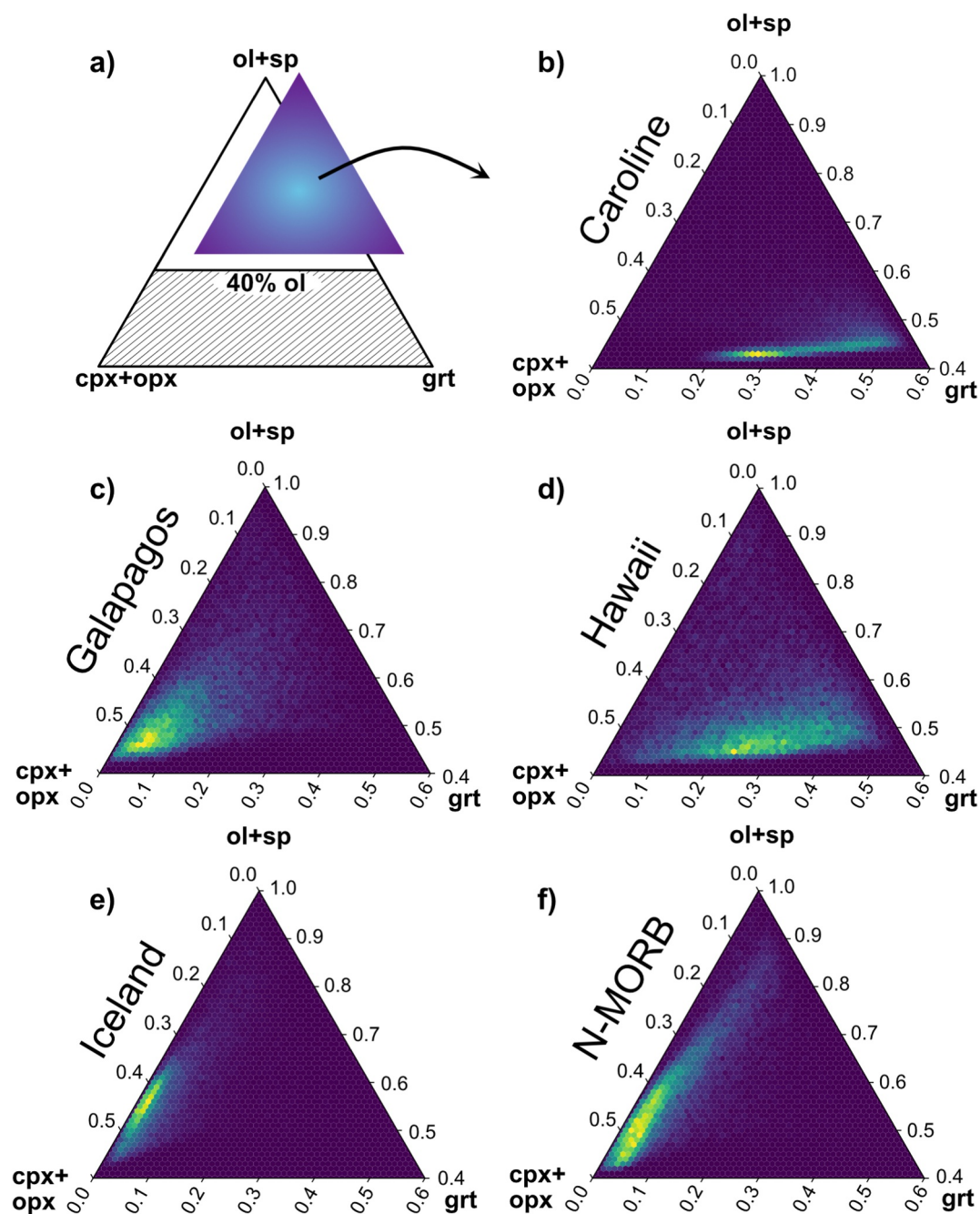


Figure 6. Density plots showing the output source mineralogies that best reproduced the average basalt elemental ratios in the ternary diagram Olivine + Spinel – Orthopyroxene + Clinopyroxene – Garnet (note truncated axes represented in panel (a)) for (b) Canaries, (c) Galapagos, (d) Hawaii, (e) Iceland, (f) average MORB (Gale et al., 2013) (see Figure S2 in Supporting Information S1 for the other islands).

Information S1. Aitchison means of mineral outputs and median melt fractions are given in Table 1. Aitchison means for all tested IOBs and the average MORB are summarized in Figure 8.

For each island, our model highlights that a large range of source mineralogies can reproduce the average basalt composition (Figure 6 and Figure S8 in Supporting Information S1). A similar observation can be made for MORB (Figure 6f). However, the basalt suites also show contrasting distributions between islands. Iceland, Meteor-Shona, Easter, and Galapagos show relatively olivine-rich mineralogy and low modal garnet. Conversely,

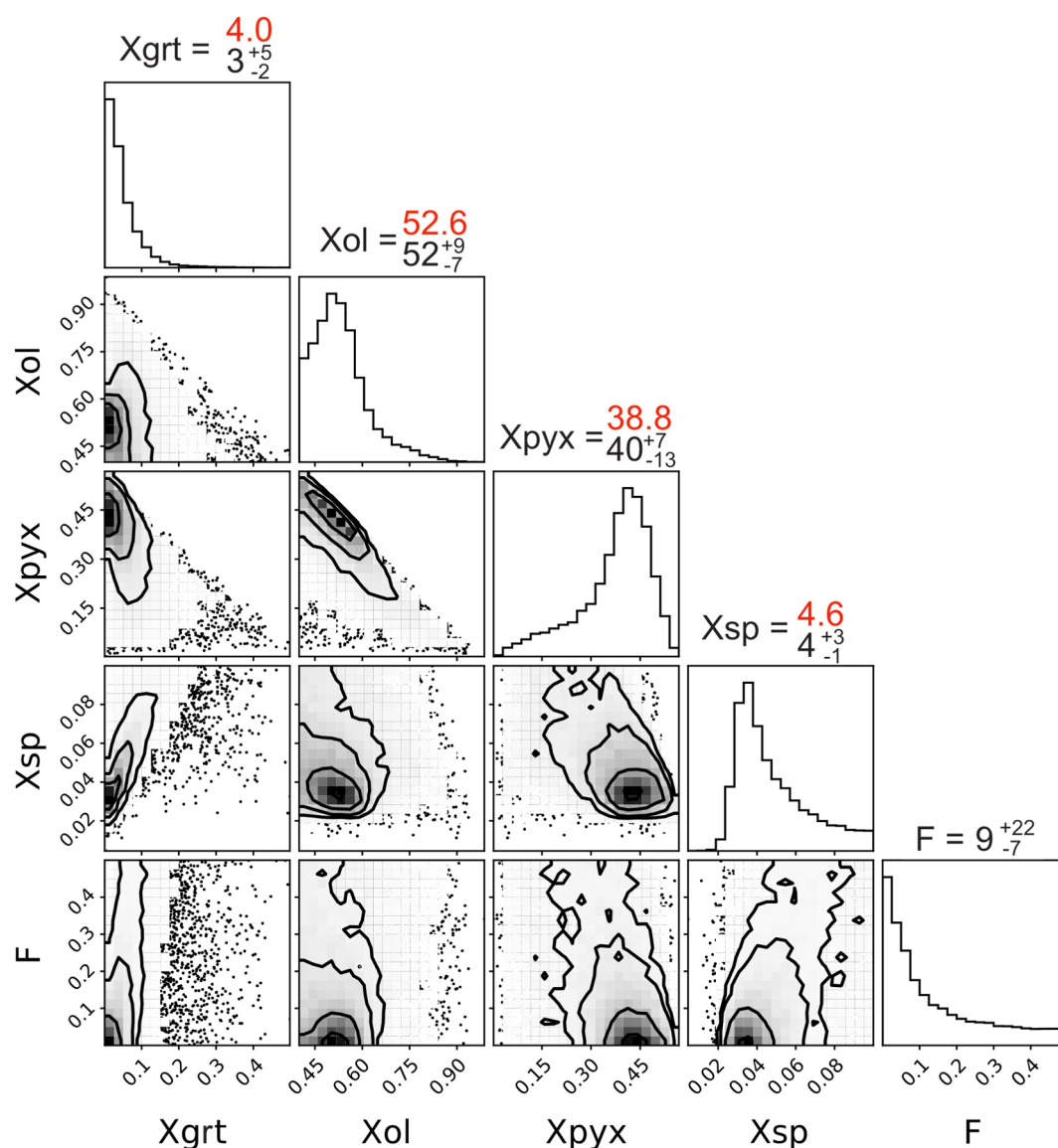


Figure 7. Example of corner plot for Iceland. The top of each column shows the posterior probability distribution for each parameter. The numbers in black represent the posterior median and the upper and lower uncertainties from the 84th and 16th percentiles, respectively. The number in red is the Aitchison mean. Output densities are plotted for each parameter pair, many showing strong correlation (e.g., Xpyx vs. Xol, Xsp vs. Xgrt). Corner plots for the other islands and MORB are presented in Figures S3–S24 in Supporting Information S1.

Tasmanid, Macdonald, Caroline, and Canary all point toward very high (>40%) modal garnet (Table 1). Hawaii predominantly shows intermediate behavior. Finally, the output mineralogy for Samoa stands out for simultaneously being very rich in pyroxenes (~47%) and poor in olivine (~41%; Figure 8). The melt fraction also varies among the OIBs, from ≤5% for Bermuda, Tristan-Gough, Meteor-Shona, and Pitcairn, to ≥10% Macdonald, Marquesas, Reunion, Samoa, and Tasmanid-Tasman Central (Table 1). The meaning of these calculated melt fractions is discussed below.

4. Discussion

Inverting elemental ratios for mantle mineralogy and melt fraction is inherently challenging due to the similar partitioning behavior between elements and coexisting mineral phases, making it difficult to uniquely attribute observed elemental signatures to a specific mineral assemblage. Despite these challenges, when applied to a

Table 1

Aitchison Means Outputs and Plume Characteristics for Each Tested Ocean Island Basalt and the Average N-MORB

OIB	Mineralogy (from Aitchison means) (%)				F (%)	V_A	T_{Pex} (Bao) ^a	T_{Pex} (Li) ^b	R_A ^c	B_{flux} (Mg/s) ^d	LAB (km) ^e	DPS ^f	Mantle components ^g					
	Ol	Pyx	Grt	Sp									P	DM	EMI	EMII	H	L
Azores	49.26	20.62	25.21	4.91	7	0.58	93	18	14.8	1	36 ± 5	0.83	X					
Bermuda	53.34	17.16	24.32	5.18	5	0.85	107			1.1	55 ± 10	0.07						
Canary	43.88	8.84	43.43	3.85	7	0.40	134		9.7	0.8	95 ± 5	0.86	X					
Cape Verde	47.95	11.78	35.63	4.64	7	0.50	50	43	15.7	1.4	95.2 ± 0.4	0.93	X					
Caroline	42.48	11.14	42.47	3.91	6	0.45	123		12.8	1.1	96.0 ± 0.1	0.86		X				
Easter	57.67	22.36	15.91	4.06	6	0.73	155		26	1.6	27.6 ± 0.8	1	X					
Galapagos	53.72	27.49	12.27	6.51	6	0.75	185	134	30.3	1.5	55 ± 5	1	X	X				
Hawaii	49.05	16.61	28.58	5.76	8	0.66	171	165	35.3	5.6	75 ± 5	1		X	X			
Iceland	52.63	38.80	3.97	4.60	9	0.43	221	180	42.9	5.2	32.3 ± 7	1		X				
Macdonald	43.01	9.77	42.00	5.23	10	0.36	254		15.8	2.6	56 ± 13	1						X
Marquesas	49.25	13.20	33.31	4.24	14	0.50	257	141	14.5	2.6	75 ± 5	0.86				X		
Meteor-Shona	60.15	19.27	13.58	7.00	4	0.76	105		12.3	0.5	<10	0.5						
Pitcairn	49.38	21.88	23.73	5.01	3	0.57	192		11.8	2.1	53 ± 1	0.92			X			
Reunion	49.05	22.64	23.53	4.78	10	0.51	125	202	15.3	1.5	70 ± 5	1						X
Samoa	40.97	47.16	11.35	0.52	15	0.09	189		35.2	1.9	93.5 ± 0.2	1				X		
Tasmanid	41.95	4.88	50.02	3.16	11	0.32	173			0.9		0.17						
Tristan-Gough	51.21	21.04	21.77	5.98	5	0.65	162	263	7	1	58.4 ± 0.4	0.71			X			
N-MORB	57.82	32.78	5.58	3.81	7	0.60	0				5.74 ± 0.27							

Note. P, PREMA; H, HIMU; L, LOND. Text in the last row (N-MORB) was italicized to demonstrate it was modeled as a reference and it is not an ocean island, unlike the other rows. ^a T_{Pex} ($T_{\text{Pex}} = T_{\text{Plume}} - T_{\text{ambient_mantle}}$) reported in Bao et al. (2022) using velocities from SEMUCB-WM1 (French & Romanowicz, 2014). ^b T_{Pex} reported in Li et al. (2025) using Al-in olivine crystallization temperatures. ^cMaximal value reported in OBID for each island, except values for Easter and Meteor-Shona (Jackson et al., 2017). ^dHotspot buoyancy fluxes reported in OBID (Harðardóttir & Jackson, 2025a, 2025b). ^eThickness of the LAB at the time of volcanism. Data are from Dasgupta et al. (2010) except for Galápagos (Gibson & Geist, 2010), Hawaii (Matthews et al., 2021), Réunion (Fontaine et al., 2015), Marquesas (Humphreys & Niu, 2009), and Meteor-Shona (assumed lower than 10 km). The lithosphere thickness for N-MORB is the thickness reported for Siqueiros in Li et al. (2025). ^fDeep Plume Score (Koppers et al., 2021): Probability, from 0 to 1, for the hotspot to be a deep plume according to the authors. ^gMantle component “clans” defined by White et al. (2025).

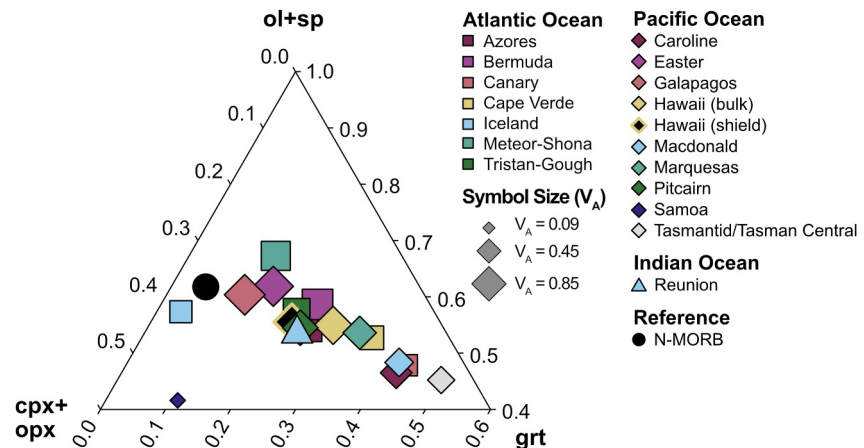


Figure 8. Aitchison means of all tested ocean island basalt suites plotted by ocean basin. Symbol size represents uncertainty in the Aitchison means, informed by the Aitchison variance (V_A ; Table 1). Axes are truncated as in Figure 6.

global suite of OIBs, BaSIC reveals a range of apparent mineralogical assemblages in the mantle sources of ocean islands. In this section, we discuss relationships between modeled source mineralogy and other characteristics of OIBs (e.g., mantle potential temperature, maximal $^3\text{He}/^4\text{He}$, LAB depth, buoyancy flux, isotopic affinity, major element composition). We then demonstrate how the source mineralogy of lavas from the recent eruptions on Iceland's Reykjanes Peninsula have shifted during the successive eruptions (2021–2024), and how they are largely distinct from older Reykjanes Peninsula eruptions.

4.1. Relationships Between Source Mineralogy and Plume Characteristics

Mantle plume theory (Morgan, 1971) has been debated since it was proposed as the process behind hotspot activity decades ago (Anderson, 2005). Since then, the nature of mantle plumes has been heavily investigated, with key divergences focusing on the source depth of plumes, their compositional heterogeneity (i.e., which mantle reservoir they are sampling), and whether hot spots are actually hot.

Several methods have been proposed to constrain the mantle potential temperature (T_p) (Cheng et al., 2024; Herzberg & Asimow, 2015; Herzberg et al., 2023; Lee et al., 2009). For OIBs specifically, mantle potential temperature compilations have been produced using both petrological (Herzberg & Gazel, 2009; Li et al., 2025; Putirka, 2008) and seismic methods (Bao et al., 2022). Using seismic velocities beneath ocean islands, Bao et al. (2022) demonstrated that while the T_p beneath plumes is generally hotter than ridges, not all hotspots are “hot.” Out of the 46 hotspots studied, they found that approximately 40% were not hot enough to facilitate active upwelling. Based on the temperature excess (T_{pex} , the difference between plume and ambient mantle temperature), they further classified hot, warm, and cold hotspots (Bao et al., 2022).

The study of OIB geochemistry has also revealed a wide range in helium ratios ($^3\text{He}/^4\text{He}$; R_A) from ocean islands, with R_A values both higher (Baffin Island; Horton et al., 2023) and lower (Tristan da Cunha; Kurz et al., 1982; Le Roex et al., 1990) than the values for MORBs (Harðardóttir & Jackson, 2025b; Horton et al., 2023; Jackson et al., 2017). Correlations between high R_A values, high buoyancy flux (Jackson et al., 2017; Putirka, 2008), and high T_p (Bao et al., 2022; Harðardóttir & Jackson, 2025b; Jackson et al., 2017) have also been demonstrated. These results have led to the hypothesis that hot mantle plumes tap a deep, hot reservoir rich in primordial helium (Harðardóttir & Jackson, 2025b; Jackson et al., 2017).

When applied to a global suite of OIBs, mineral outputs from BaSIC show correlations with several of these mantle plume characteristics (Figure 9). Results are also displayed as a correlation matrix in Figure S25 in Supporting Information S1. Here we summarize our main observations:

1. The thickness of the lithosphere (Figure 9a) and the proximity of the island to a spreading center (Figure 9h) are the two parameters that exert the most control on the apparent mineralogy of the source. Hotspots that are isolated and located above thick lithosphere produce mineral assemblages that are poorer in olivine (and spinel) than those located close to a spreading center and above thinner lithosphere. Except for Samoa, they are also significantly richer in garnet.
2. OIBs with source assemblages that have higher modal garnet also tend to have relatively low $^3\text{He}/^4\text{He}$ ratios (Figure 9c).
3. Except for the Caroline islands, OIBs with deeper lithosphere asthenosphere boundaries (LAB) also exhibit geochemical evidence of a recycled component in their source (EM1, EM2, HIMU or LOND, Figure 9e; White et al., 2025). OIBs with olivine-rich source mineralogies primarily record DM or PREMA signatures.
4. The inverted source mineralogies show no apparent correlation with the potential temperature (Figure 9b) or the buoyancy flux (Figure 9d).
5. The calculated apparent melt fraction (Figure 9f) is broadly correlated with the mantle potential temperature (Figure 9b), and, to a lesser extent, with the lithospheric thickness (Figure 9a).

4.2. Implications for the Nature of the Mantle Sources of Oceanic Islands

The average mineral assemblage displayed for each island in Figure 8 is unlikely to directly represent the bulk mineralogy of the mantle beneath each island. In fact, numerous pieces of evidence show that the mantle is heterogeneous and contains at least two, and likely multiple types of lithologies (e.g., Anderson, 2006; Ballmer et al., 2011; Brown et al., 2020; Elkins & Lambart, 2024; Grand et al., 1997; Herzberg, 2011; Hirschmann &

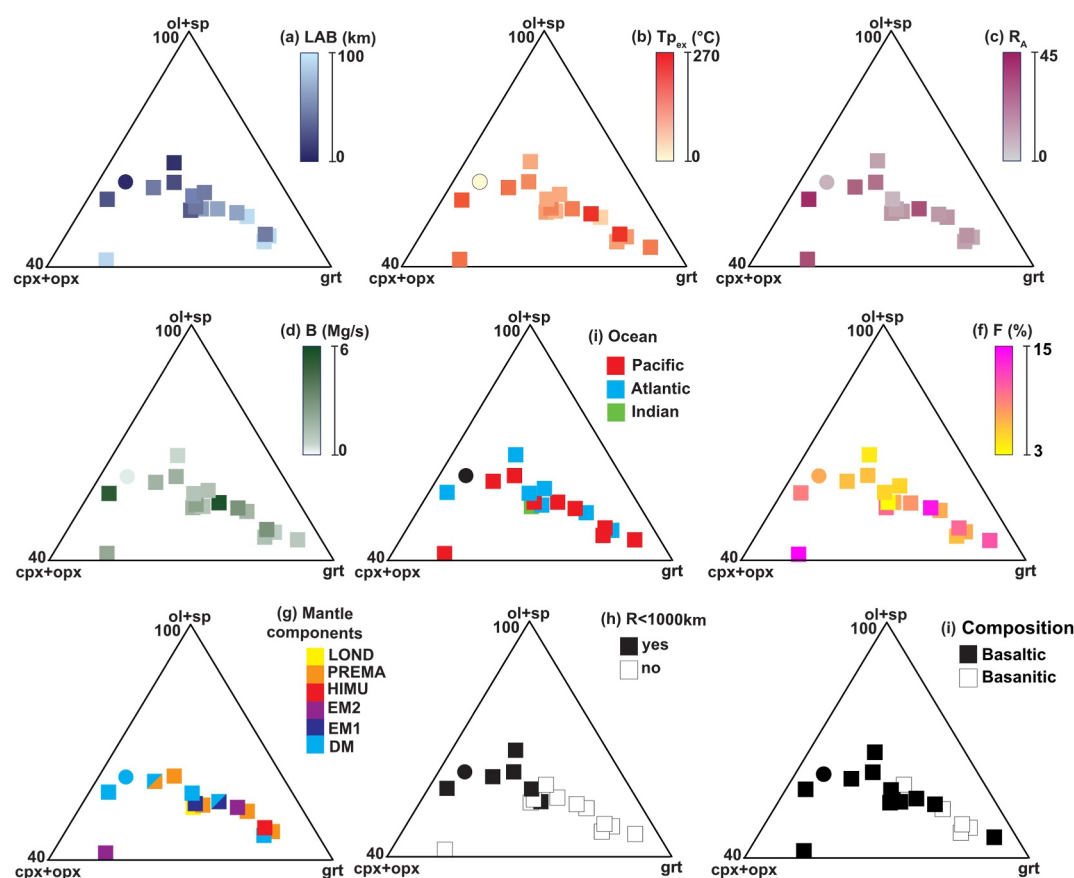


Figure 9. Aitchison mean outputs from BaSIC plotted in the ternary diagram olivine + spinel – cpx + opx – garnet with (a) LAB thickness, (b) T_{pex} from Bao et al. (2022), (c) maximum RA, (d) buoyancy flux, (e) location, (f) modeled apparent melt fraction, (g) mantle component clans, (h) presence of ridge segments within 1000 km radius of the island, (i) composition. See also Figure S25 in Supporting Information S1 for a correlation matrix, and Table 1 for the complete list of references.

Stolper, 1996; Hole, 2018; Stracke et al., 2019) and that each island likely samples a different mantle source (Bao et al., 2022).

Many pyroxenites have lower solidus temperatures than peridotites (e.g., Hirschmann & Stolper, 1996; Lambart et al., 2016). In addition, pyroxenites also have significantly higher melt productivities, and can contribute disproportionately to the melt relative to their abundance in the source (e.g., Hirschmann & Stolper, 1996; Lambart et al., 2009; Mallik et al., 2021; Phipps Morgan, 2001; Stolper & Asimow, 2007). Early melting of pyroxenite also results in a local decrease of the temperature of the host peridotite (e.g., Brunelli et al., 2018; Phipps Morgan, 2001; Weatherley & Katz, 2016). Decreasing T_p or increasing the thickness of the lithosphere results in further increase of the pyroxenite contribution to the melt (e.g., Mallik et al., 2021; Sims et al., 2013; Stracke et al., 2000). For a given T_p , however, because pyroxenitic material is usually denser than peridotite, a mantle plume can only accommodate a small fraction of pyroxenite while sustaining active upwelling (Bao et al., 2022; Brown & Leshner, 2014). Buoyancy may nonetheless be preserved (or even enhanced) if the mantle plume also contains relatively refractory, and therefore less dense, material (Béguelin et al., 2025; Sanfilippo et al., 2024; Shorttle et al., 2014; Stracke et al., 2022). Additionally, because refractory components will undergo less melting, their presence in the source can also amplify the relative contribution of fertile lithologies to the aggregate melt (Figure S26 in Supporting Information S1).

Considering a multilithological mantle, the apparent mineralogy detected by BaSIC will consequently reflect a combination of the tectonic setting (namely, the mantle temperature and thickness of the lithosphere) and of the contributions of each lithology present in the source. The strong tectonic control on the mineralogical output is

exemplified by the comparison between Hawaii and Iceland, the two islands that present the highest buoyancy flux of all OIBs in our data set (Figure 9b). The presence of refractory material in the mantle source has been recently suggested to explain the high buoyancy for both islands (Béguelin et al., 2025; Sanfilippo et al., 2024; Shorttle et al., 2014) and is supported by geochemical and isotopic evidence pointing toward a depleted component in their respective mantle sources (e.g., DeFelice et al., 2019; Fitton et al., 1997; Harrison et al., 2020; Thirlwall, 1995). The two islands are also both characterized as hot spots by Bao et al. (2022) with similar potential temperatures, and record very high $^3\text{He}/^4\text{He}$ ratios (Figure 9c). Despite these similarities, BaSIC infers contrasted source mineralogies between the two islands (Figure 8). Tectonic setting is the main difference between the islands: Hawaii is an isolated hotspot on a relatively thick lithosphere (Table 1), whereas Iceland sits on the mid-Atlantic ridge. The contrasted mineralogical results between the two islands are consistent with melting of a heterogeneous source under lithospheres of differing thickness. In Hawaii, under thicker lithosphere, the portion of the melting column within the peridotite-melting interval is shorter, resulting in a higher apparent contribution of fertile lithologies with higher melt productivities (e.g., Mallik et al., 2021). With a thinner lithosphere, like beneath Iceland, the larger melting column effectively dilutes the pyroxenite signature. Several studies have supported the presence of lithological heterogeneity in the mantle sources of many of the oceanic island basalt suites discussed here (e.g., Ballmer et al., 2011; Barker et al., 2014; Béguelin et al., 2025; Brown et al., 2020; Chauvel et al., 2008; DeFelice et al., 2019; Gurenko et al., 2010; Koornneef et al., 2012; Nebel et al., 2019; Sanfilippo et al., 2024; Shorttle & MacLennan, 2011; Sobolev et al., 2005, 2007; Soderman et al., 2023; Wang et al., 2021). Though clinopyroxene is generally considered as the primary component of pyroxenite, due to the difficulty in parsing types of pyroxene in models such as BaSIC, we interpret source mineralogies that are rich in garnet as likely reflecting the presence of a pyroxenitic lithology. In fact, some of the apparent mineralogies reported in Table 1 show garnet fractions >0.4 . To reach 40% garnet in the subsolidus assemblage, Thermocalc calculations performed on lherzolite (KLB-1) show that it requires a pressure of 130 kbar (see Figure 6 in Shorttle et al. (2014)), well above the pressure estimated for the peridotite solidus to cross the adiabatic path, even for the maximal T_p observed in this data set (~ 50 kbar, Lambart et al., 2016). Many of the islands with garnet-rich mineralogies also have relatively low $^3\text{He}/^4\text{He}$ values, suggesting a recycled origin for the lithological heterogeneity. This is contrasted by the high $^3\text{He}/^4\text{He}$ ratios recorded in Hawaii and Iceland, which may suggest that the fertile material present in these two plumes (and responsible for these ratios) differs in origin from the fertile lithologies present beneath the other islands, or that an additional mantle component is carrying the high R_A signature.

Similarly, the apparent melt fraction does not reflect the average melt fraction of the mantle source, but rather the average melt fraction of each lithology weighted by its contribution to the aggregated magmas. To illustrate this, it is interesting to compare the outputs for Pitcairn and Réunion. Both islands show similar apparent mineralogy (Figure 8), but their apparent melt fractions are contrasted, with 3% for Pitcairn and 10% for Réunion (Table 1). Pitcairn sat on a thinner lithosphere at the time of volcanism (Dasgupta et al., 2010; Fontaine et al., 2015) and records a higher temperature anomaly (Bao et al., 2022) and a higher mantle buoyancy flux (Harðardóttir & Jackson, 2025b) than Réunion (Table 1). So, if the source was the same for both islands, we would expect Pitcairn to record a higher melt fraction than Réunion. In the context of a bilithological mantle, the colder mantle and thicker lithosphere for Réunion imply that melting initiates at shallower depth and that, for a given fraction (and composition) of pyroxenite in the source, the contribution of the pyroxenite-derived melt will be higher than for Pitcairn. This is consistent with the higher apparent degree of melting for Réunion (Stolper & Asimow, 2007). To explain their similar apparent mineralogy, this also suggests that the fraction of pyroxenite in the source of Réunion must be smaller.

Remarkably, the apparent source mineralogy appears to be decoupled from the isotopic affinity of the mantle source (Figure 9g). For instance, Samoa and Marquesas both carry a strong isotopic signature for the mantle component EMII, but show very contrasted mineralogies, with Samoa mantle source being significantly richer in pyroxenes (Figure 8, Table 1). A similar observation can be made for Iceland and the Caroline islands that are both considered as having a primarily DM isotopic signature (White et al., 2025) while their apparent source mineralogies fall at the opposite ends of the trend produced by most islands (Figure 8). Inversely, the Caroline islands, Canary, and Macdonald all show similar apparent source mineralogies while presenting contrasted isotopic affinities. Jackson and Dasgupta (2008) demonstrated that HIMU hotspot lavas tend to have lower SiO_2 and higher FeO and $\text{CaO}/\text{Al}_2\text{O}_3$ than lavas with other affinities. They interpreted this observation as pointing toward a carbonated mantle (peridotite + eclogite) source for the HIMU lavas. This differs from our results,

which suggest that the apparent source mineralogy is not correlated with isotopic affinity (Figure 9g). As mentioned in Jackson and Dasgupta (2008), the alkalic HIMU hotspots (Cook-Australis and St. Helena in their data set) exhibiting low SiO_2 and high FeO^T and $\text{CaO}/\text{Al}_2\text{O}_3$ produce small magma fluxes. We propose that their distinct major element compositions also reflect a tectonic control. If the mantle source is lithologically heterogeneous (e.g., containing both peridotite and pyroxenite), a small magma flux will result in oversampling the pyroxenitic lithology and will result in a high apparent degree of melting and average pressure of melting, despite low magmatic productivity (Stolper & Asimow, 2007).

Further, Stracke et al. (2022) note that although volcanic rocks of the Southwest Pacific form an isotopically distinct group, there is no simple relationship between the isotopic affinity, inferred mantle upwelling structures from seismology (Obayashi et al., 2016), or temporal evolution of volcanism (Chauvel et al., 2012; Koppers et al., 2008) in the region. Hence, the lithospheric structure may also have an important control on volcanic activity, supporting models involving secondary solid-state mechanical mixing (Herzberg & Gazel, 2009; Lambart, 2017; Shorttle & MacLennan, 2011) and melt-rock reactions (Lambart et al., 2012; Pilet et al., 2008, 2011; Sobolev et al., 2005, 2007), in contrast to models favoring a more direct contribution of recycled lithologies to the compositions of erupted melts (e.g., Dasgupta et al., 2010; Hauri, 1996). Finally, we acknowledge that these interpretations can also be complicated by other processes occurring during the melting process, such as incomplete melt aggregation (e.g., Koornneef et al., 2012), various mixing processes (Rudge et al., 2013), or melt reequilibration and chromatographic effects during transport (Kogiso, Hirschmann, & Reiners, 2004; Liang et al., 2011; Navon & Stolper, 1987).

Many of the islands producing garnet-rich mineralogies also appear to be sub-alkaline (Figure 9i). Melting of metasomatic amphibole veins in the lithosphere has been suggested as a driver in the genesis of oceanic alkaline basalts (Pilet, 2015; Pilet et al., 2008, 2011). As mentioned in Section 2.1.1, ignoring the potential presence of residual amphibole in the mantle source may result in an overestimation of the garnet and olivine fraction. We tested the inclusion of amphibole in our model by running additional inversions considering a mantle source composed of olivine, pyroxene (using the partition coefficients of cpx), amphibole, garnet, and spinel. We ran this model for Cape Verde, Macdonald (both showing an average basanitic composition; Figures S27–S29 in Supporting Information S1), and Marquesas (subalkaline composition; Figures S27 and S30 in Supporting Information S1) islands. In these three cases, the Aitchison means are very similar to those obtained assuming an anhydrous (i.e., amphibole-free) mantle source, and the calculated fraction of amphibole is systematically lower than 5%. These results do not necessarily mean that amphibole did not contribute to the melting process. However, if present, amphibole is likely a minor residual phase in the mantle source. In fact, when present, amphibole is usually the first phase to be consumed during the melting process (e.g., Médard et al., 2006; Pilet et al., 2008). Alternatively, Cunningham, Lambart, Guo, et al. (2026) demonstrated that the circulation of anhydrous silicate melt at the base of a 60 km thick lithosphere can result in the formation of clinopyroxene-garnet veins. Remobilization of such veins during melting would lead to a stronger residual signal from garnet. This garnet signal would persist to high melt fractions, long past the point where amphibole would be exhausted (e.g., Lambart et al., 2013; Pertermann & Hirschmann, 2003). Hence, the high modal garnet fractions obtained for several islands might reflect these metasomatic events.

4.3. Case Study: The Recent Eruptions on Iceland's Reykjanes Peninsula

The beginning of the Fagradalsfjall Fires in 2021 signaled the end of the nearly 800-year period of volcanic quiescence on Iceland's Reykjanes Peninsula (Sæmundsson et al., 2020). Since then, activity on Reykjanes Peninsula has continued, with concurrent eruptions at Fagradalsfjall in 2022 and 2023 (the Fagradalsfjall Fires). At the end of 2023, a new eruption began at Svartseng (Sundhnúksfígur eruption), ~5 km west of the Fagradalsfjall Fires, which continued until May 2024 (Matthews et al., 2024; Troll et al., 2024). Trace element and radiogenic isotopic compositions of these recently erupted basalts suggest they are compositionally distinct from greater Iceland basalts (Halldórsson et al., 2022; Marshall et al., 2024; Troll et al., 2024).

We apply BaSIC to basalts from the recent (Bindeman et al., 2022; Troll et al., 2024) and older (GEOROC) Reykjanes Peninsula eruptions (Figure 10; Table S2 in Supporting Information S1). Output mineralogies from basalts of the first recent eruption (2021, Geldingadalir) plot near basalts from older Reykjanes Peninsula and Western Volcanic Zone eruptions. However, the fraction of pyroxene increases in the following eruptions (2022–2023). Despite the large uncertainty for the calculated melt fraction, results also suggest the recent

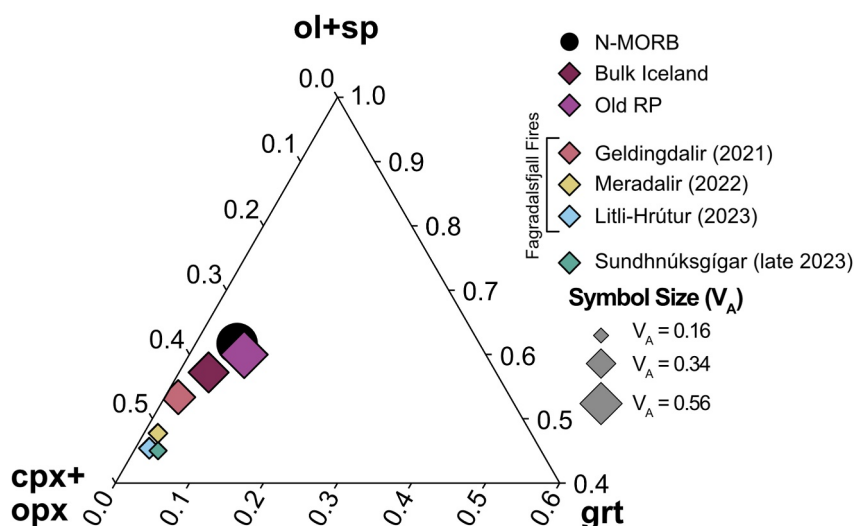


Figure 10. Aitchison means for each recent (2021–2023) Reykjanes Peninsula eruption. Symbol size represents uncertainty in the Aitchison means, informed by the Aitchison variance (V_A ; Table S2 in Supporting Information S1).

eruptions are derived from a significantly lower melt fraction than that inferred for the older Reykjanes Peninsula basalts (Figure 11, Table S2 in Supporting Information S1).

Day et al. (2024) showed that the rhenium-osmium isotopic compositions of the recent lavas require some degree of crustal contribution. However, both Fagradalsfjall lavas (2021–2023) and Sundhnúksgrígar lavas (2023–2024) have similar $\delta^{18}\text{O}$ values, with values expected for a normal upper mantle (Bindeman et al., 2022; Troll et al., 2024). Because crustal contamination does not significantly affect the elemental ratios used in BaSIC (see Section 2.1), it cannot be responsible for the observed trend in mineralogy space. We also discard the potential effect of differentiation as the ratios used in BaSIC are also relatively insensitive to differentiation (see Section 2.1), and the composition of the lavas only show a slight decrease of MgO content with time (from 9.0 to 8.2 wt%).

The 2022–2023 Fagradalsfjall Fires have distinctively more enriched Sr and Nd isotopic compositions than the older Reykjanes Fires (800–1240 AD; Peate et al., 2009; Matthews et al., 2024). The lavas are also more homogeneous than the lavas erupted during the 2021 Fagradalsfjall eruption. Though all recent eruptions fall on the same La/Sm versus Nb/Zr trend, the 2022–2023 Fagradalsfjall fires show the highest La/Sm and Nb/Zr ratios and are less scattered (Figure 12a). La/Sm is a proxy for the degree of melting (e.g., Stracke & Bourdon, 2009), and Nb/Zr has been used as a tracer of lithological heterogeneity in Iceland (Shorttle et al., 2014). Coupled with the results obtained with BaSIC, these trace element observations are consistent with the 2022–2023 eruptions representing a relatively low degree of melting of an enriched reservoir (Halldórsson et al., 2022; Troll et al., 2024).

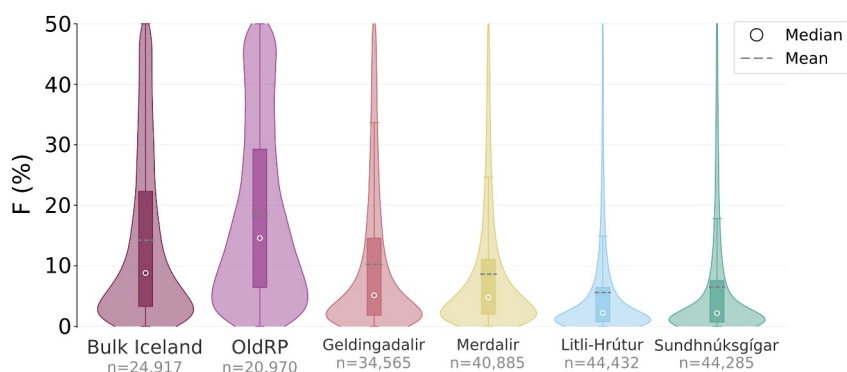


Figure 11. Violin plots of modeled melt fraction from BaSIC, showing the melt fraction decreases with concurrent eruptions.

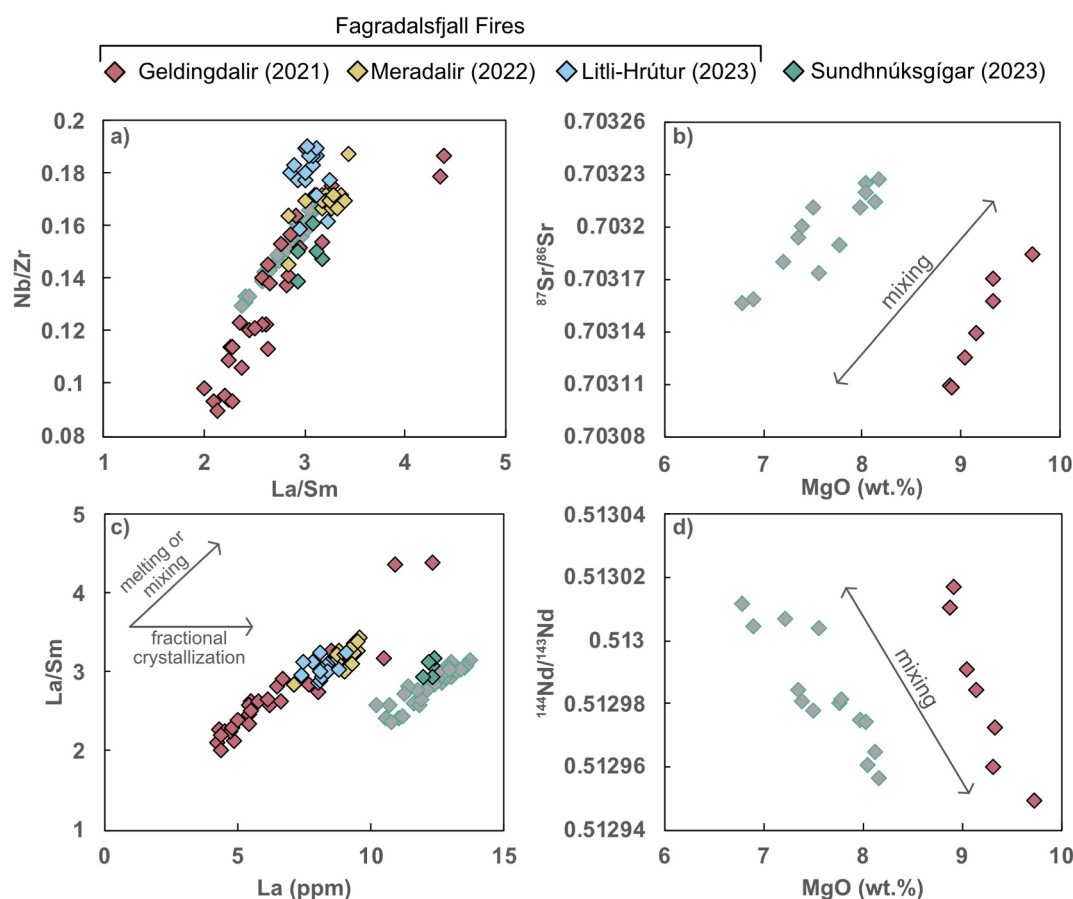


Figure 12. Compositions of the lavas reported by Troll et al. (2024) for the recent (2021–2023) Reykjanes Peninsula eruptions. We also plotted the compositions reported by Matthews et al. (2024) for the Sundhnúksíggar eruption (gray symbols).

In the model results shown in Figure 10, only geochemical data from the first day of the most recent eruption (Sundhnúksíggar) are included (Troll et al., 2024). Matthews et al. (2024) subsequently reported major, trace, and isotope compositions on lavas during the various stages of the eruption between December 2023 and May 2024. Unfortunately, we are unable to use these data in BaSIC, as they do not include Ga concentrations. However, it is still informative to look at the evolution of the Sundhnúksíggar lava composition with time. First, the comparison of trace element concentrations between the different eruptions argues in favor of a higher degree of fractional crystallization for the Sundhnúksíggar eruption (Figure 12c), consistent with the lower MgO contents. Second, both the 2021 Fagradalsfjall eruption and the Sundhnúksíggar eruption show correlation between MgO content and Sr and Nd isotopic ratios (Figures 12b and 12d), consistent with sampling peridotitic (lower Sr ratio) and pyroxenitic (higher Sr ratio) components. Neave et al. (2019) showed that during differentiation, when melts are derived from different lithologies, the melt derived from more fusible lithologies tends to be oversampled in the aggregated magmas due to their lower liquidus temperature; further fractionation should result in a progressive increase of the bias toward fusible lithologies. We observe the opposite trend, however, with the most primitive magmas within a single eruption showing the most isotopically enriched compositions (Figures 12b and 12d). In addition, while the lavas became progressively more enriched during the 2021 Fagradalsfjall eruption (Hall-dórsson et al., 2022), the opposite evolution was observed during the Sundhnúksíggar eruption (Matthews et al., 2024). Hence, the trends observed for the 2021 Fagradalsfjall and the Sundhnúksíggar eruptions are unlikely to reflect fractionation trends of a reservoir with melts derived from various mantle components (Matthews et al., 2024), but are rather a mixing trend between an evolved melt that is isotopically depleted and a more primitive melt that is isotopically enriched, stored in two distinct and isolated reservoirs. This is consistent with BaSIC's results (Figure 10), where the 2021 eruption represents an intermediate composition between the

depleted component, predominantly sampled by the Old RP and Bulk Iceland data sets, and the enriched component, which is heavily sampled by the 2022 and 2023 eruptions.

We note that the lithological makeup of the mantle source for both reservoirs does not necessarily need to significantly differ. In fact, both BaSIC calculations and trace element concentrations suggest that the 2022–2023 lavas are produced by a lower degree of melting (Figures 11 and 12). Low degree melts will preferentially sample the most fusible lithology (pyroxenite) and hence, can explain the apparent increase of pyroxene (Figure 10), the higher Nb/Zr ratios (Figure 12a), and the more enriched isotopic ratios. The offset in Sr isotopic composition between the 2021 Fagradalsfjall eruption and the Sundhnúksgrígar eruption observed by Matthews et al. (2024) may either be explained by a slight difference in isotopic composition of the enriched components between the two eruptions (Matthews et al., 2024) or may result from various magma mixing processes (Rudge et al., 2013).

5. Conclusions

We introduce BaSIC (Basalt Source Investigation and Characterization), a Bayesian inversion model that uses six elemental ratios to constrain the source mineralogy and melt fraction of basalts. By combining a modal aggregated fractional melting framework with nested sampling, BaSIC efficiently explores a high-dimensional mineralogical parameter space and returns a range of plausible source mineralogies rather than a single solution.

When applied to a global suite of OIBs, BaSIC reveals substantial variability in inferred source mineralogies across ocean islands. The inferred mineralogies show no systematic dependence on mantle potential temperature, isotopic affinities, or buoyancy flux. On the contrary, the dominant control appears tectonic, where islands that are isolated from spreading centers and situated above thick lithosphere yield source assemblages enriched in garnet and depleted in olivine. In contrast, islands that are near ridges or above thin lithosphere show more olivine-rich, garnet-poor assemblages. This pattern is consistent with a model in which pyroxenitic lithologies are present throughout the mantle source of OIBs, but their apparent contribution to the aggregate melt depends strongly on lithospheric thickness, which controls the length of the melting column, and hence the relative contributions of the pyroxenite and peridotite components.

The comparison between Hawaii and Iceland is particularly instructive. Despite sharing similarly high potential temperatures, high buoyancy fluxes, and elevated $^3\text{He}/^4\text{He}$ ratios, the two islands yield different source mineralogies. We interpret this contrast as a direct expression of their differing lithospheric settings, with the thick lithosphere beneath Hawaii amplifying the pyroxenite signal that is diluted by a longer melting column beneath Iceland. BaSIC also reveals a decoupling between isotopic affinity and bulk mineralogy. Iceland and the Caroline Islands, for example, are both considered to have a primarily DM isotopic signature yet show contrasting source mineralogies, indicating secondary processes such as mechanical mixing and melt-rock reactions are important processes in generating OIB diversity, rather than a direct contribution of isotopically distinct source components.

BaSIC's capacity to track source evolution within a single volcanic system is demonstrated using the Reykjanes Peninsula as a case study. BaSIC detects a systematic increase in the pyroxene mode and a decrease in melt fraction across the 2021–2023 Fagradalsfjall eruptions, consistent with progressive sampling of an enriched, more fusible reservoir at lower melt fractions. These findings align with isotopic and trace element evidence for increasing source enrichment through the eruption sequence and are consistent with a two-reservoir model in which a depleted peridotitic component and an enriched pyroxenitic component are sampled in varying proportions.

BaSIC is openly available to the community as a Python package (see Open Research Statement) and is intended for use across a range of intraplate and ridge settings. It is best suited for comparative studies across multiple basalt suites or for tracking source evolution through time, where relative changes in mineralogy are more interpretable than absolute modal fractions. Model outputs should be interpreted with the understanding that BaSIC inverts for the bulk apparent mineralogy of the melting region rather than individual lithologies, and that geologic non-uniqueness, which is inherent to any type of inversion, is captured in the reported Aitchison variance.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

All files required to run BaSIC are available at Cunningham, Lambart, and Brown (2026). The repository contains three files: a model script that contains the most complex code, a driver script that is meant to be edited and will call and run the model script, and a ReadMe file with extensive instructions for using the model and a troubleshooting guide.

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