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An ancient, deep, and ultra-depleted mantle domain beneath the northern Atlantic Ocean

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Abstract

Mantle differentiation produces geochemically enriched crust and complementary depleted mantle residues. While crustal recycling over supercontinent cycles has shaped a first-order east-west mantle compositional dichotomy between an enriched African mantle domain and a depleted Pacific mantle domain, the nature, distribution and origin of depleted mantle remain poorly known. Here we report the presence of an ultra-depleted mantle domain beneath the northern Atlantic Ocean, characterized by highly depleted elemental compositions and radiogenic Hf isotopes. We develop a statistical mantle model to reconstruct distinct mantle depletion histories, which can also explain the Hf-isotope discrepancy between oceanic basalts and peridotites. Model results indicate that the ultra-depleted mantle (UDM) domain records ancient high-degree melt

extraction at depths greater than those of normal mid-ocean ridge melting. In contrast, a depleted Pacific mantle domain reflects a progressive melt extraction that mostly occurred at shallow depths, compatible with those of mid-ocean-ridges. We propose that distinct melt-depletion histories contribute to mantle's compositional structure, in which the UDM domain beneath the northern Atlantic Ocean may represent residues of ancient mantle plumes that have been preserved in the lower mantle.

Introduction

Earth's differentiation and long-term geodynamic evolution have produced a heterogeneous convective mantle, recorded by oceanic basalts and peridotites. Recent statistical analyses of global basalt geochemistry revealed an east-west mantle dichotomy, expressed as a relatively enriched African mantle domain and a more depleted Pacific mantle domain^{1,2}. This dichotomy is geographically coupled with the two large low-shear-wave-velocity provinces (LLSVPs), indicating that enriched crustal materials have been preferentially transported into the African mantle domain by long-lived subduction during Pangea assembly and along the circum-Pacific subduction girdle. While the effects of crustal recycling on enriched mantle heterogeneities have been extensively investigated¹⁻⁴, the nature and origin of depleted mantle remain poorly documented and understood.

The depleted upper mantle has traditionally been treated as a comparatively homogeneous reservoir produced by crust-mantle differentiation, but this paradigm has increasingly been challenged by evidence for the presence of ultra-depleted mantle components⁵⁻¹⁰. For example, abyssal peridotites from the Gakkel and Southwest Indian ridges can have abnormally low Al_2O_3 , unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ and radiogenic $^{176}\text{Hf}/^{177}\text{Hf}$, recording ancient, refractory and ultra-depleted mantle parcels^{5,6,8,11}. Such observations have led to the hypothesis that ultra-depleted mantle may be widespread or even volumetrically significant within the asthenosphere⁹. Yet abyssal peridotites are sparsely sampled and unevenly distributed, limiting constraints on the global distribution of ultra-depleted mantle. By contrast, oceanic basalts broadly sample mantle beneath global mid-ocean ridge system and ocean islands/plateaus, but how they

record depleted mantle heterogeneities remains enigmatic⁹.

Here we integrate global geochemical observations of oceanic basalts and peridotites to reconstruct a global mantle depletion framework that reflects the mantle's complex depletion history. We report the presence of an ultra-depleted mantle (UDM) domain beneath the northern Atlantic Ocean that is distinguishable from the depleted Pacific mantle domain by coupled elemental and Nd-Hf isotopic systematics, and interpret this UDM domain as an ancient one in the lower mantle that was brought to the surface through Cenozoic plume activity.

An ultra-depleted mantle domain beneath the northern Atlantic Ocean

We compile isotopic and elemental data of MORBs, plume-related ocean island basalts (OIBs) and oceanic plateau basalts (OPBs) along the global mid-ocean ridges and near-ridge hot spots (Fig. 1). MORBs primarily sample the shallow asthenosphere, whereas plume-related OIBs and OPBs commonly incorporate deeper mantle components^{2,12}. Given a heterogeneous mantle composed of peridotites and pyroxenites, MORBs and plume-related OIBs and OPBs sample different portions of mantle components^{13,14}. While Sr, Nd and Pb isotopic signatures of oceanic basalts can be easily overwhelmed by enriched mantle components, Hf isotopes, owing to the moderate incompatibility of Lu and Hf, are more sensitive indexes of depleted and ultra-depleted mantle components^{15,16}.

The Nd-Hf isotopic patterns of MORBs define depleted, ultra-depleted and enriched mantle domains (Figs. 1 and 2), signatures of which are also observed in near-ridge OIBs and OPBs (Supplementary Fig. 1) from the Pacific, northern Atlantic, southern Atlantic and Indian Oceans. Pacific MORBs define a depleted mantle domain, characterized by high ϵ_{Nd} and radiogenic ϵ_{Hf} , but predominantly negative $\Delta\epsilon_{\text{Hf}}$ ($\Delta\epsilon_{\text{Hf}} = \epsilon_{\text{Hf}} - 1.55\epsilon_{\text{Nd}} - 1.21$) relative to the terrestrial array¹⁷ (Fig. 2a). By contrast, African MORBs display more heterogeneous Nd-Hf isotopic signatures that can be subdivided into two domains: i) MORBs from the northern Atlantic Ocean (from the Azores–Iceland region northward to the Knipovich ridge) (Fig. 1) exhibit high ϵ_{Nd} and positive $\Delta\epsilon_{\text{Hf}}$ that clearly plot above the terrestrial array, defining a UDM domain; ii) MORBs

85 from the southern Atlantic and Indian Oceans exhibit overall low ϵNd and dispersed
86 ϵHf around the terrestrial array, defining a relatively enriched mantle domain. In
87 addition, MORBs from the central Atlantic Ocean exhibit highly variable isotopic
88 signatures transitional between the abovementioned two mantle domains^{4,10}, and will
89 not be considered further in this study.

90 While the African mantle domain overall exhibits more enriched and hydrous
91 signatures relative to the Pacific mantle domain^{1,2,4}, our analysis shows that the UDM
92 domain beneath the northern Atlantic Ocean displays distinct signals from the enriched
93 mantle domain beneath the southern Atlantic and the Indian Ocean (Figs. 2b–d). The
94 UDM domain exhibits concentrated and lower $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ relative to
95 the PREMA^{18–20}, whereas the enriched mantle domain has dispersed and higher
96 $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{208}\text{Pb}^*/^{206}\text{Pb}^*$. Moreover, the UDM domain exhibits higher Nb/Nb^*
97 ($\text{Nb}/\text{Nb}^* = \text{Nb}_\text{N}/(\text{Th}_\text{N}^{0.5} \times \text{La}_\text{N}^{0.5})$, where ‘N’ denotes primitive mantle normalization²¹)
98 and lower Th/U relative to the enriched mantle domain, although they consistently have
99 higher Ba/Th than the Pacific mantle domain² (Supplementary Fig. 2).

100 In addition to isotopic compositions, the UDM domain displays systematically
101 depleted major-element and incompatible-element compositions. At comparable high
102 MgO contents ($\text{MgO} > 8.0$ wt. %), MORBs from the UDM domain exhibit lower TiO_2
103 and Na_2O relative to those from the depleted and enriched mantle domains (Figs. 3a–
104 b), consistent with melts derived from clinopyroxene-poor peridotites²². Incompatible
105 element ratios such as Zr/Y also support an ultra-depleted mantle source (Figs. 3c–d),
106 although the slightly elevated Nb/Y ratios indicate a possible contribution of recycled
107 oceanic crust²³. While depletion in elements can be explained by high-temperature
108 mantle melting associated with mantle plumes²⁴, these ultra-depleted characteristics
109 persist even at distances > 1000 km from plume centers (Supplementary Figs. 3–4),
110 indicating that they reflect a widespread UDM source⁸ rather than localized high-
111 temperature melting. Together, these coupled isotopic and elemental characteristics
112 suggest the existence of a UDM domain beneath the northern Atlantic Ocean.

Global mantle depletion modelling based on statistical analyses

Because MORBs and abyssal peridotites are commonly regarded as complementary melts and melting residues derived from the same mantle domain, they have traditionally been expected to preserve consistent isotopic characteristics²⁵. Recent studies, however, demonstrate that abyssal peridotites can have extremely high ϵ_{Hf} relative to MORBs of the same region^{6,8,9,26,27}, here referred to as the Hf isotopic paradox. One explanation is that ultra-depleted peridotites with high ϵ_{Hf} are too refractory to melt efficiently beneath mid-ocean ridges^{5,6}. However, experimental constraints suggest that clinopyroxene-poor peridotites can melt under normal ridge conditions, albeit producing lower melt fractions than fertile peridotites²² (Supplementary Fig. 5). In addition, some peridotites with extremely high ϵ_{Hf} retain relatively fusible major-element compositions^{8,9,28}. Together, these observations indicate that refractory behavior alone is insufficient to explain the systematic Hf isotopic differences between abyssal peridotites and MORBs.

An alternative explanation is that melts from isotopically ultra-depleted peridotites are strongly depleted in incompatible elements, which are subsequently overprinted by enriched mantle melts during magma ascent and mixing^{7,9,26}. As such, MORBs with highly depleted incompatible-element compositions should tend towards high $\Delta\epsilon_{\text{Hf}}$ with increasing ϵ_{Nd} . This prediction, however, is inconsistent with observations of the Garrett transform in the Pacific Ocean, where some of the most incompatible-element-depleted MORBs²⁹ exhibit extremely high ϵ_{Nd} but negative $\Delta\epsilon_{\text{Hf}}$, following the depleted Pacific mantle trend³⁰ (Fig. 2a). The ultra-depleted mantle endmember developed by Stracke et al.⁶ satisfactorily explains Nd-Hf isotope compositions of peridotites, but it does not readily account for the systematic Hf isotopic offset between the depleted and ultra-depleted mantle domains recorded by oceanic basalts, unless variably depleted endmembers are invoked^{9,16}.

To reproduce the links between ultra-depleted peridotites and basalts, we utilize Monte Carlo approaches to model mantle composition under variable depletion extents and ages (Methods, Supplementary Figs. 6–10). Published global abyssal peridotite

results are taken as representing the entire range and frequency distribution of the depleted mantle assemblage, from which variable depletion extents were generated using observations of abyssal peridotites corroborated by a dynamic melting model (Supplementary Fig. 8). For a depleted MORB mantle (DMM)³¹-like mantle assemblage formed at 800 Ma within the spinel peridotite field, the modelled Nd–Hf isotopic compositions overlap those observed in abyssal peridotites (Supplementary Fig. 9), although the modelled proportion of extremely radiogenic- ϵ Hf compositions is lower than observed. It is noteworthy that extremely radiogenic- ϵ Hf peridotites are more likely to be sampled when a modelled mantle assemblage has greater melt-depleted extents and/or older depletion ages, consistent with previous work^{6,8}. In addition, modelled mantle assemblages generated within the garnet peridotite field develop higher Lu/Hf ratios and more time-integrated radiogenic ϵ Hf than those formed in the spinel peridotite field (Supplementary Fig. 9), providing a potential mechanism for generating the distinction between the ultra-depleted and depleted mantle domains (Figs. 2a and 3c).

Distinct depletion histories between the ultra-depleted and depleted mantle domains

Two-component mantle mixing models reveal that the ultra-depleted and depleted mantle domains reflect distinct mantle depletion processes (Supplementary Fig. 6) (Methods). We first consider a shallow-depletion model (Scenario I), in which varying mantle assemblage models formed exclusively within the spinel peridotite field³² are mixed with recycled oceanic crust, both formed at 800 Ma as a starting age. The modelled mixtures reproduce the ϵ Nd range of global MORB mantle, but yield slightly lower $\Delta\epsilon$ Hf that is below the terrestrial array (Supplementary Fig. 11). Increasing depletion degrees of the modelled mantle assemblage would shift these mixtures toward higher ϵ Nd, yet $\Delta\epsilon$ Hf remains lower, with only sparse modelled data extending to higher $\Delta\epsilon$ Hf. This demonstrates that normal mid-ocean-ridge depletion alone is unlikely to account for the UDM domain.

Alternatively, scenario II assumes that mantle depletion occurs at greater depths

than normal mid-ocean-ridge melting, involving variable proportions of spinel- and garnet-peridotite residues³³ at 800 Ma. With increasing mantle depletion, modelled mixtures evolve towards higher ϵNd and higher $\Delta\epsilon\text{Hf}$, producing a substantially improved match to the UDM domain (Fig. 4). For an ultra-depleted mantle assemblage with a mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.32 (more depleted than the Depleted-DMM)³¹ the modelled isotopic field closely reproduces the observed UDM domain (Fig. 4d). This agreement remains robust even with higher proportions of garnet-peridotite residues (Supplementary Fig. 12) or older depletion ages (e.g., 1.5 Ga; Supplementary Fig. 13). These results indicate that the UDM domain requires an ancient, ultra-depleted mantle assemblage formed by high-degree melting at depths greater than normal mid-ocean-ridge melting.

By contrast, neither Scenario I nor II reproduces the high ϵNd and low $\Delta\epsilon\text{Hf}$ endmember of the depleted Pacific mantle domain (Supplementary Fig. 11 and Fig. 4). Instead, scenario III, which involves ancient oceanic lithosphere (oceanic crust and lithospheric mantle)³⁴ followed by continuous shallow melt extraction, provides a viable explanation. Recycled oceanic crust can generate low $\Delta\epsilon\text{Hf}$ values below the terrestrial array³⁵, whereas subsequent shallow depletion shifts the mantle towards higher ϵNd . While this model employs a simplified two-stage oceanic lithosphere depletion (oceanic lithosphere formed at 1.5 Ga followed by melt extraction at 800 Ma), the resulting mixtures successfully reproduce the combination of high ϵNd and low $\Delta\epsilon\text{Hf}$ for the depleted Pacific mantle domain (Supplementary Fig. 14).

Together, these results indicate that the UDM and depleted mantle domains record distinct mantle depletion histories in Earth's history. Our modelling results suggest that the UDM domain in the northern Atlantic Ocean likely experienced an ancient (≥ 800 Ma) mantle depletion event at greater-than-mid-ocean-ridge melting depth and high-temperature³⁶. Such conditions can be readily provided by large-scale ancient plume-related melting events (Fig. 5a). Given that the modelled depletion ages represent minimum estimates, even older Hadean-Archean mantle residues^{37,38} appear to be reasonable candidates³⁹. By contrast, the depleted Pacific mantle domain records

recycling of ancient oceanic lithosphere, followed by progressive melt extraction at shallow mid-ocean-ridge melting depths. This is consistent with the formation of the paleo-Pacific since the break-up of the supercontinent Rodinia^{1,2,40}.

Geodynamic implications of the ultra-depleted mantle

The existence of UDM components beneath the northern Atlantic Ocean, which are composed of substantial clinopyroxene-poor peridotites³⁶, has important implications for mantle geodynamics^{7,8}. If UDM components were stored primarily in the upper mantle, their positive buoyancy relative to more fertile mantle lithologies would have favored long-term accumulation beneath ridges, and should have been recorded by earlier northern Atlantic MORBs. However, MORBs formed during the opening of the northern Atlantic Ocean (e.g., lavas from Goban Spur and Hatton Bank) exhibit high ϵNd but negative $\Delta\epsilon\text{Hf}$ ([Supplementary Fig. 15](#)), distinct from observations of the present-day UDM domain⁴¹. Alternatively, this UDM signature since the eruption of the North Atlantic Igneous Province might have originated from delamination of sub-continental lithospheric mantle during continental break-up³⁰. However, as discussed above, the post-55 Ma appearance of the UDM signature ([Supplementary Fig. 15](#)) similarly argues against this model.

Here we hypothesize that the UDM components beneath the northern Atlantic Ocean were formed by plume-related high-degree melting at ca. 800 Ma or older. The ultra-depleted mantle lithospheric blocks were recycled to the lower mantle ([Fig. 5a](#)), and remained there^{7,8} until Cenozoic mantle plumes brought them up and produced basalts with the reported UDM signatures ([Fig. 5b](#)). Studies suggested that clinopyroxene-poor peridotites can indeed be long-lived in the lower mantle^{7,42,43}. The deep plume origin is spatiotemporally consistent with the appearance of ultra-depleted mantle signatures^{7,8,14,39} ([Supplementary Fig. 15](#)). In addition, the geographic distribution of the ultra-depleted mantle domain overlaps regions influenced by the (proto-)Iceland and the (proto-)Azores plumes⁴⁴, where positive residual topography exists⁴⁵. UDM components from the lower mantle may also contribute to other plumes (e.g., the Marion Rise, Sierra Leone and Hawaii)^{10,46,47}, although their signatures are

often obscured by enriched components in plume-derived basalts. If this is indeed the case, ancient depleted components in the deep mantle must be taken into account in models of Earth's geodynamic evolution and mantle compositional heterogeneity^{40,42,48}. The presence of such ancient depleted components in the deep mantle may also help explain the anomalous buoyancy and upward flux of mantle plumes^{8,14}.

Methods

Compilation and filtering of geochemical data

Geochemical data used in this study ([Supplementary Table 1](#)) were compiled from published literature, supplemented by published datasets from Yang et al.⁴ and Harðardóttir et al.⁵¹. We focus exclusively on Hf isotope data of oceanic basalts measured by multi-collector inductively coupled plasma mass spectrometry (MC-ICP-MS), whereas those analyzed by thermal ionization mass spectrometry (TIMS) are not included. MORB samples with MgO < 5.0 wt % were excluded to minimize the effects of magmatic differentiation.

Monte Carlo modelling of mantle compositions

To parameterize mantle depletion, we fitted a Johnson-SU distribution to the ¹⁴⁷Sm/¹⁴⁴Nd ratios of abyssal peridotites compiled in [Supplementary Table 2](#). The fitted parameters were $a = -7.4960$, $b = 1.7667$, $\text{loc} = -1.8316$ and $\text{scale} = 0.0208$. The fitted distribution had a mean ¹⁴⁷Sm/¹⁴⁴Nd ratio of approximately 0.37 and represented the most depleted end-member assemblage considered in our model. To reduce the effects of complex melt-rock interaction and metasomatic overprinting, only abyssal peridotites with ¹⁴⁷Sm/¹⁴⁴Nd > 0.2 were considered in this study. Alternative truncation thresholds, including the ¹⁴⁷Sm/¹⁴⁴Nd_{PM} = 0.196²¹ were also tested, which do not affect interpretations in this study.

To generate mantle assemblages with variable depletion extents, we first truncated the fitted Johnson-SU distribution by excluding the ¹⁴⁷Sm/¹⁴⁴Nd < 0.2 tail and renormalized the remaining probability density. The truncated distribution was then transformed toward ¹⁴⁷Sm/¹⁴⁴Nd = 0.20 using a linear compression:

$$X_{\text{new}} = 0.20 + k (X_{\text{old}} - 0.20),$$

where X_{old} represents values sampled from the truncated Johnson-SU distribution and k is the compression factor. This procedure preserves the empirical shape of the abyssal-peridotite distribution while systematically reducing its mean $^{147}\text{Sm}/^{144}\text{Nd}$. For each target assemblage, k was chosen such that the conditional mean of $^{147}\text{Sm}/^{144}\text{Nd}$ within the interval $0.2 < ^{147}\text{Sm}/^{144}\text{Nd} < 1.0$ matched a prescribed value. In this study, we considered target means of 0.24, 0.26, 0.28 and 0.32 to represent mantle assemblages with progressively increasing depletion extents (Supplementary Fig. 8). For each mantle assemblage, we used dynamic melting models to calculate corresponding Nd, Hf and $^{176}\text{Lu}/^{177}\text{Hf}$ compositions for residues formed in the spinel- and garnet-peridotite fields (Supplementary Table 3). These modelled elemental correlations allow each Monte Carlo simulation of $^{147}\text{Sm}/^{144}\text{Nd}$ to be translated into corresponding Sm, Nd, Lu and Hf, then the isotopic compositions of mantle assemblages at given depletion ages could be achieved.

As an illustrative case, we generated elemental and isotopic compositions of a DMM-like mantle assemblage, which was extracted from a starting mantle source at 800 Ma. This starting mantle source has an Nd-Hf isotopic evolution similar to that of PREMA (Supplementary Fig. 7), and is used here as an isotopic approximation of a broadly primitive mantle reservoir^{18–20}. Nd-Hf isotopic evolution of a PREMA-like source lies within errors of a depleted mantle reservoir^{35,52}. The modelled Nd-Hf isotopic compositions of mantle assemblages broadly reproduce the observed isotopic range of the abyssal peridotites (Supplementary Figs. 9a–b). Alternative source compositions, such as chondritic mantle, were also tested but produced isotopic compositions less consistent with the observed data fields (Supplementary Fig. 10). This is partly because a PREMA-like composition is a more plausible mantle source for oceanic basalts⁵³, which provides minimum estimates of depletion ages and extents. All Monte Carlo simulations were implemented using the open-source programming language Python (version 3.11).

Monte Carlo simulation of two-component mantle mixtures

We used Monte Carlo sampling to simulate two-component mixtures comprising 95–99% mantle assemblage and a 1–5% recycled oceanic crust. For a given depletion extent and age, isotopic compositions of a mantle assemblage were generated by combining variable proportions of mantle residues formed in the spinel- and garnet-peridotite fields from a PREMA-like source^{19,20,35,53}. Isotopic compositions of the recycled oceanic crust were modelled by assuming N-MORB⁵⁴ was extracted from a PREMA-like source at a specified age, similar to the approach of Chauvel et al.³⁵.

Data availability

All data used in this study can be found in [Supplementary Tables 1–4](#).

Code availability

All the code are available upon request to the corresponding author.

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Author contributions:

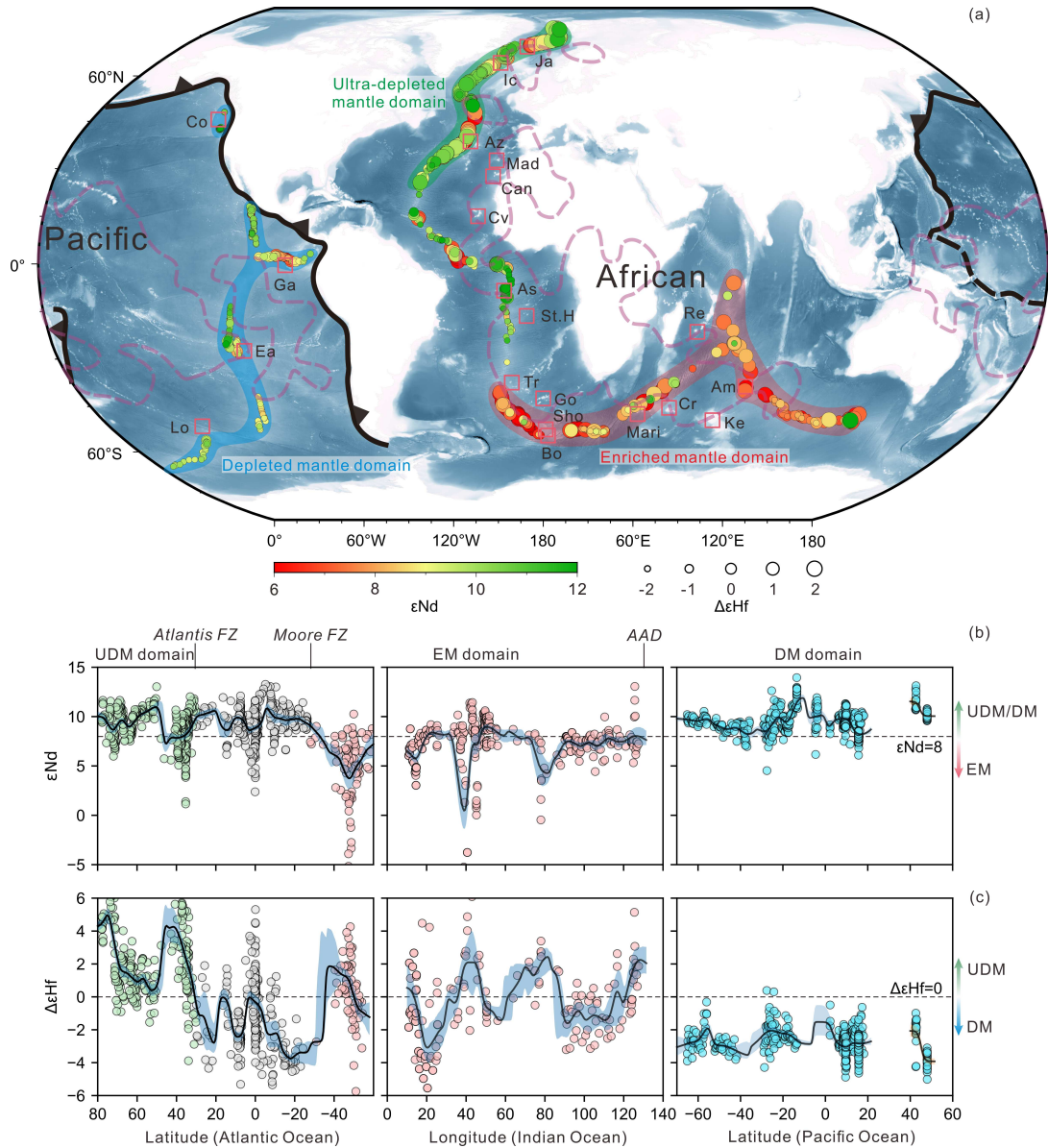
F.X. and Z.X.L. conceived and designed the study. F.X. compiled the data, developed the model and prepared the figures. F.X. drafted the manuscript. P.A.B.S., A.S., X.B.Z.,

440 L.S.D. and Z.X.L. contributed to data interpretation and critically revised the
441 manuscript. All authors contributed to discussions.

442 **Competing interests**

443 The authors declare no competing interests.

444



446

447 **Fig. 1 ϵ_{Nd} and $\Delta\epsilon_{\text{Hf}}$ values of global mid-ocean-ridge basalts (MORBs).** a,
 448 Distribution of mid-ocean-ridge basalts, with their ϵ_{Nd} and $\Delta\epsilon_{\text{Hf}}$ values shown. $\Delta\epsilon_{\text{Hf}}$
 449 $= \epsilon_{\text{Hf}} - 1.55\epsilon_{\text{Nd}} - 1.21^{17}$. Based on ϵ_{Nd} and $\Delta\epsilon_{\text{Hf}}$, we identify three mantle domains:
 450 a depleted mantle (DM) domain beneath the Pacific Ocean; an ultra-depleted mantle
 451 (UDM) domain beneath the northern Atlantic Ocean; an enriched mantle (EM) domain
 452 beneath the southern Atlantic and the Indian oceans. The purple dashed lines denote the
 453 $-1\% \text{ dln}V_{\text{S}}$ contour at 2900 km depth from the SEMUCB-WM1 model⁴⁹. The plot was
 454 produced using the open-source library PyGMT. Hotspot abbreviations: As, Ascension;

Az, Azores; Bo, Bouvet; Can, Canary; Cr, Crozet; Cv, Cape Verde; Ea, Easter; Ga, Galápagos; Go, Gough; Ic, Iceland; Ja, Jan Mayen; Ke, Kerguelen; Lo, Louisville; Mad, Madeira; Re, Réunion; St. H, Saint Helena; Tr, Tristan. Geochemical data of oceanic basalts are listed in [Supplementary Table 1. b,c](#), Isotopic variations of MORBs from the Atlantic, Indian and Pacific Oceans. MORBs from the northern Atlantic Ocean generally exhibit high ϵNd ($\epsilon\text{Nd} > 8$) and positive $\Delta\epsilon\text{Hf}$ ($\Delta\epsilon\text{Hf} > 0$), whereas those from the Pacific Ocean exhibit high ϵNd ($\epsilon\text{Nd} > 8$) but negative $\Delta\epsilon\text{Hf}$ ($\Delta\epsilon\text{Hf} < 0$), corresponding to ultra-depleted and depleted mantle domains, respectively. Black curves show Gaussian-kernel regressions of MORB data, and shaded envelopes indicate 95% bootstrap confidence intervals.

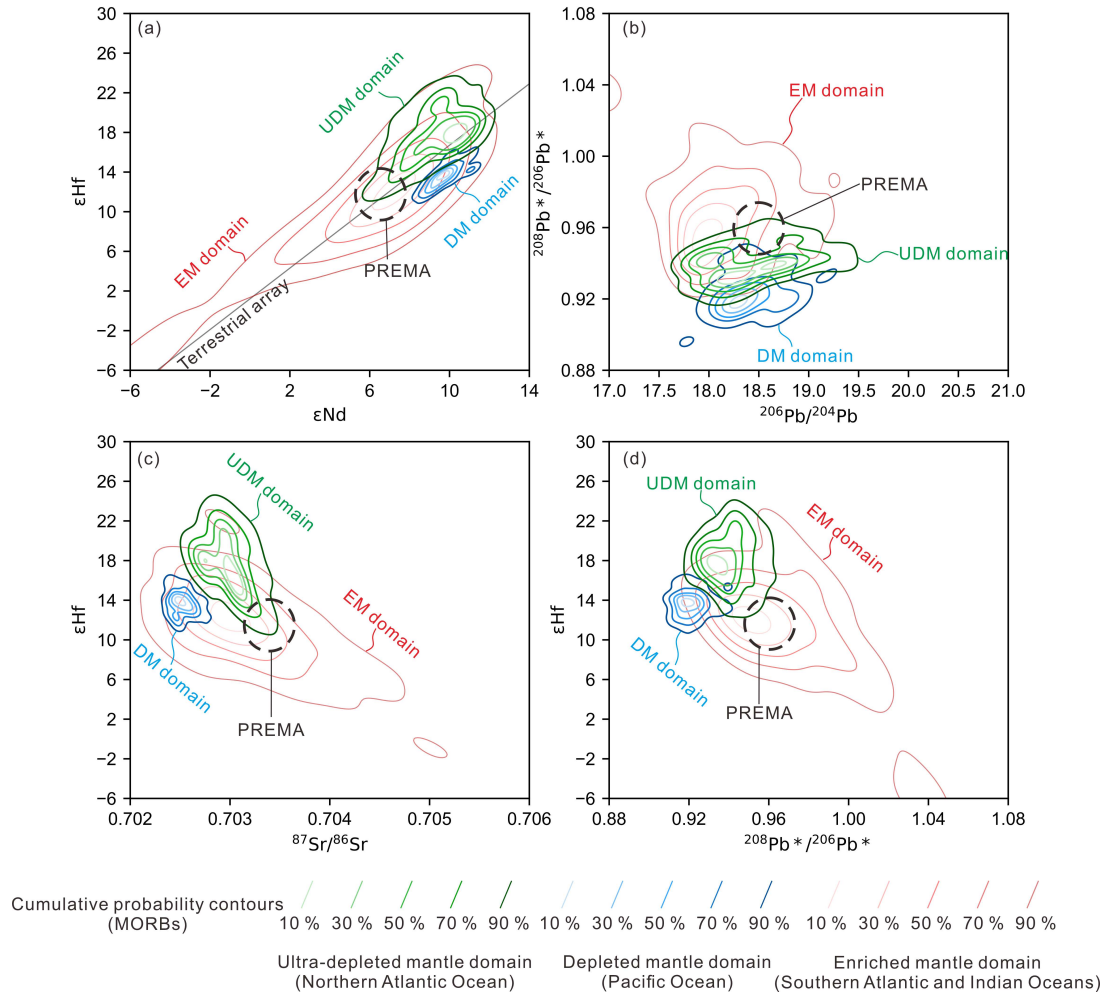


Fig. 2 Isotopic compositions of MORBs from depleted, ultra-depleted and enriched mantle domains. **a**, ϵ_{Hf} versus ϵ_{Nd} . Terrestrial array is from Vervoort et al.¹⁷. **b**, $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ versus $^{206}\text{Pb}/^{204}\text{Pb}$. $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ represents radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ of samples relative to chondrites.⁵⁰ **c**, ϵ_{Hf} versus $^{87}\text{Sr}/^{86}\text{Sr}$. **d**, ϵ_{Hf} versus $^{208}\text{Pb}^*/^{206}\text{Pb}^*$. Isotopic data of the PREMA are from Jackson and Carlson¹⁹ and Giuliani et al.²⁰. Contours denote kernel-density estimates of MORB data from the northern Atlantic Ocean, Pacific Ocean and southern Atlantic and Indian Oceans. Geochemical data of MORBs are listed in [Supplementary Table 1](#).

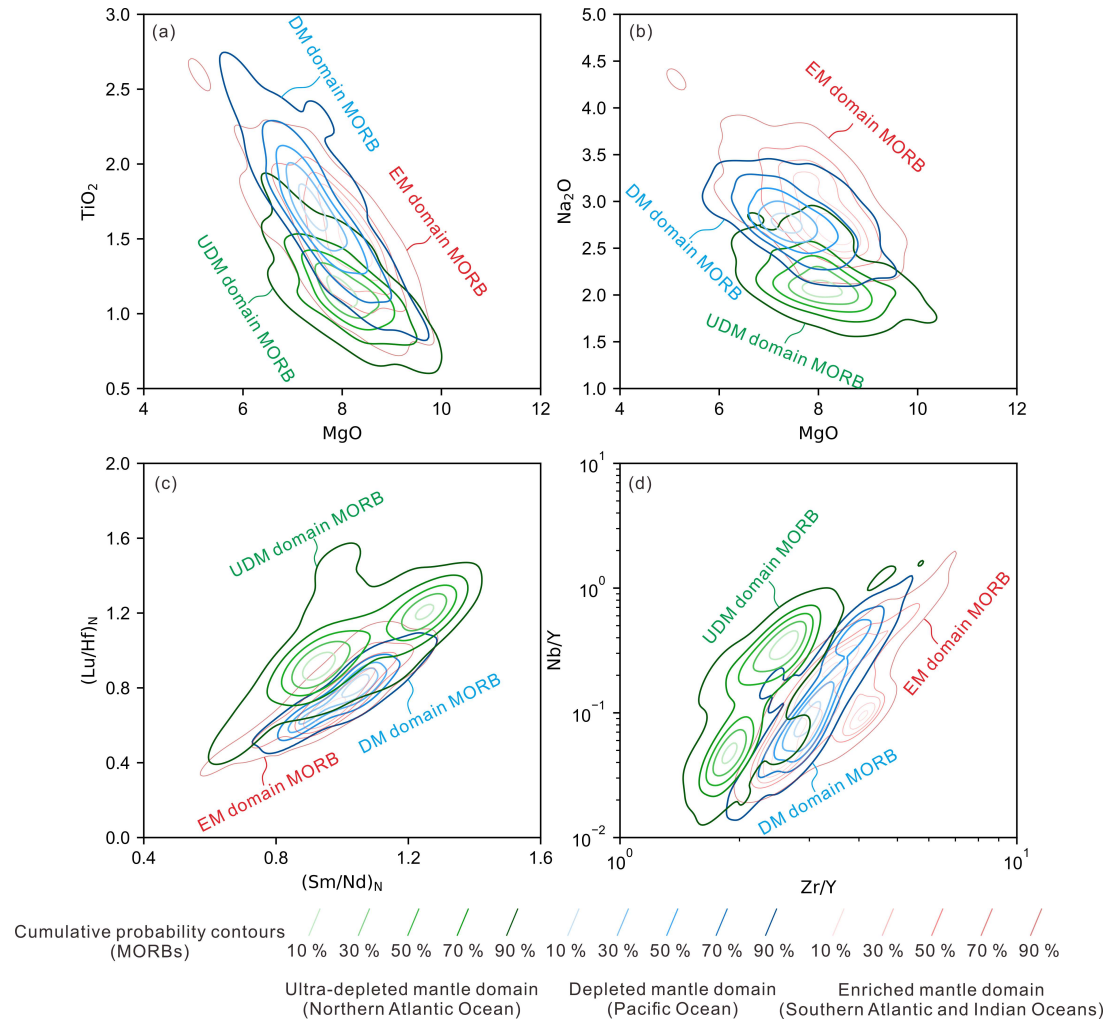


Fig. 3 Major- and trace-element compositions of MORBs from depleted, ultra-depleted and enriched mantle domains. a, TiO_2 versus MgO . b, Na_2O versus MgO . c, $(\text{Lu}/\text{Hf})_N$ versus $(\text{Sm}/\text{Nd})_N$. 'N' denotes primitive mantle normalization²¹. d, Nb/Y versus Zr/Y . Contours denote kernel-density estimates of MORB data from the northern Atlantic Ocean, the Pacific Ocean, and the southern Atlantic and Indian oceans. Geochemical data of MORBs are listed in [Supplementary Table 1](#).

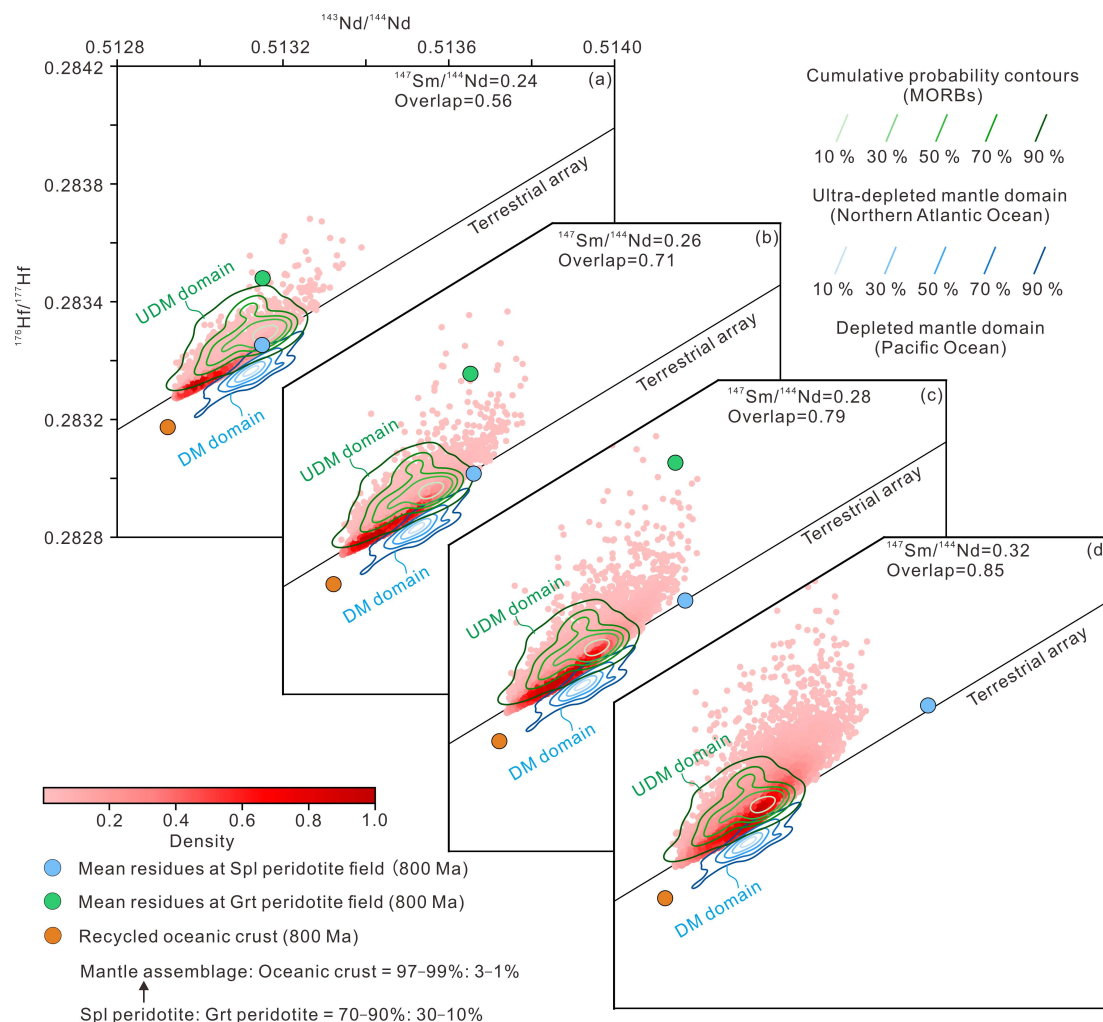
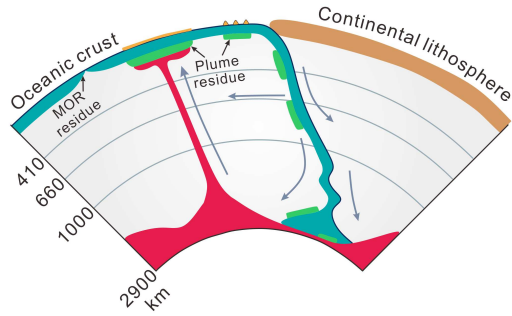


Fig. 4 Monte Carlo simulation of isotopic compositions of two-component (mantle assemblage and recycled oceanic crust) mantle mixtures. a–d, Monte Carlo results for mantle assemblages with mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.24, 0.26, 0.28 and 0.32, respectively, representing progressively greater depletion. The model workflow is shown in [Supplementary Fig. 6](#). Mantle assemblages formed at 800 Ma comprise variable proportions of spinel- and garnet-peridotite residues (spinel peridotite: garnet peridotite= 90–70%:10–30%), and are mixed with 1–3% recycled oceanic crust component of the same age. Simulated compositions are colored by relative point density. Reported overlap values quantify the areal overlap between the modelled distributions and the observed ultra-depleted mantle (UDM) field. Blue and green circles indicate mean isotopic compositions of spinel- and garnet-field residues, respectively, whereas orange circles denote recycled oceanic crust. Blue and green contours represent the observed DM and UDM domains, respectively. The black line represents the terrestrial array¹⁷.

(a) ≥ 800 Ma: formation of ancient ultra-depleted mantle
Shallow mid-ocean ridge vs. deep plume mantle melting



(b) Present-day mantle domains

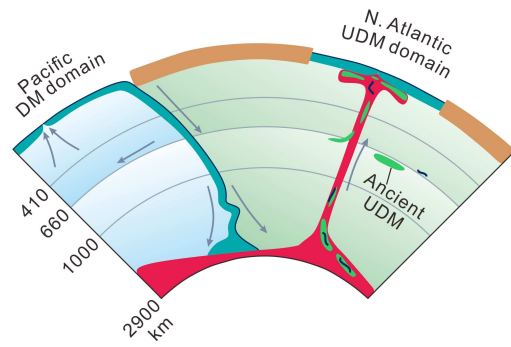
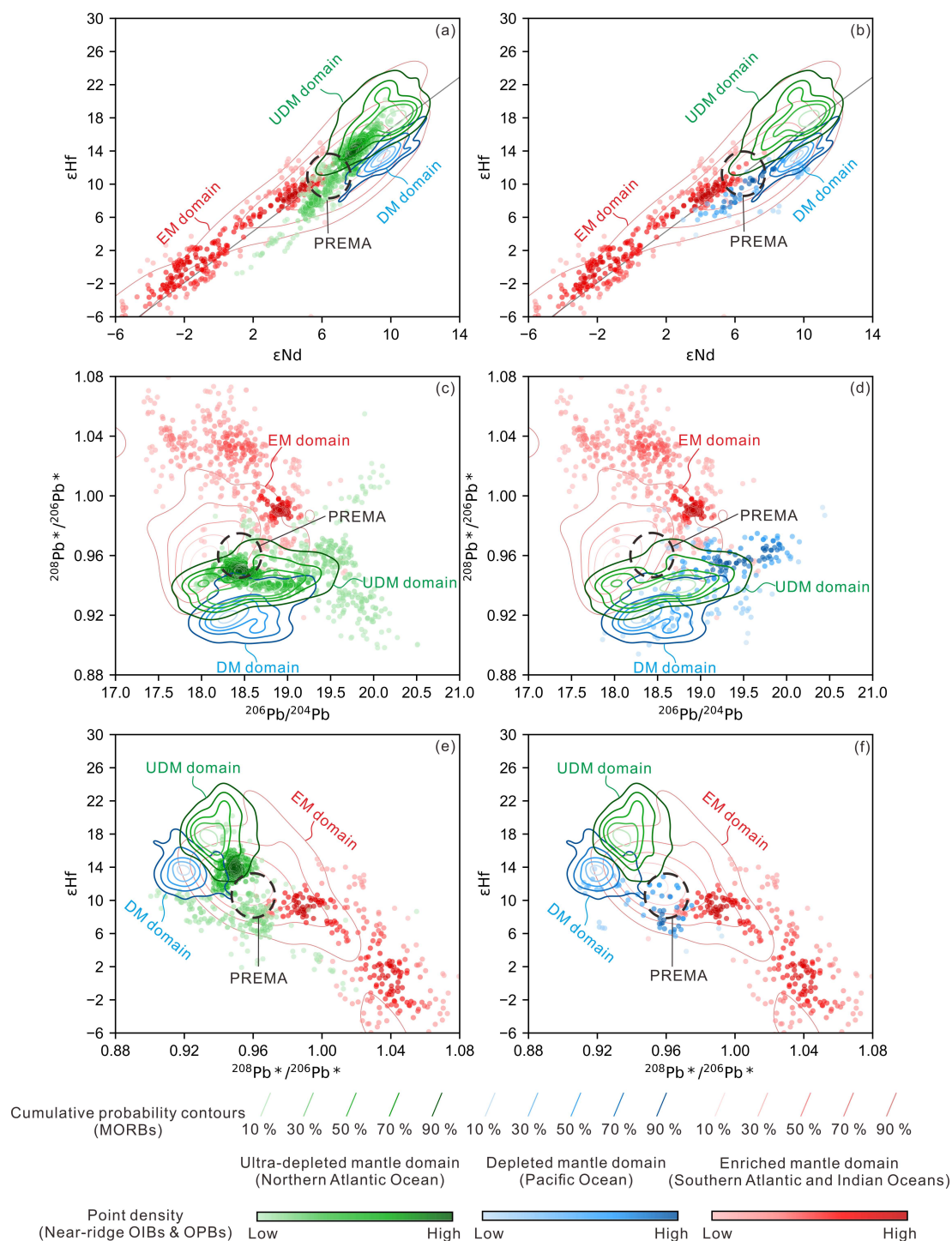
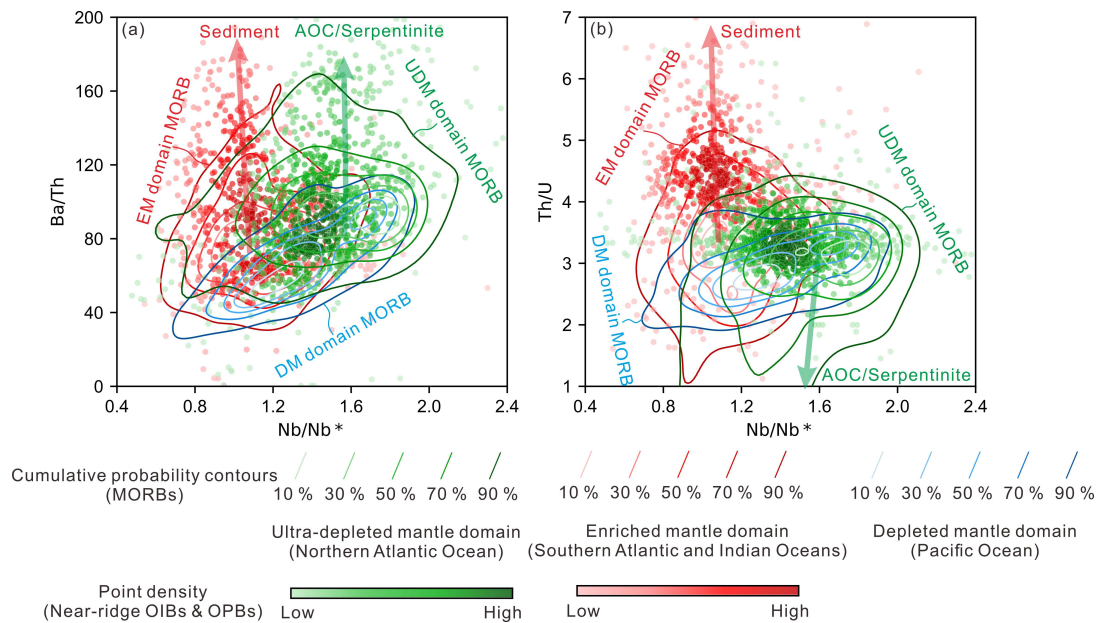


Fig. 5 Schematic for possible origins of depleted and ultra-depleted mantle domains. a, Mantle melting under plume environment prior to 800 Ma produces ultra-depleted mantle residues, whereas that under mid-ocean-ridge environment produces depleted mantle residue. **b,** The UDM domain beneath the northern Atlantic Ocean records ancient ultra-depleted mantle residues, stored in the lower mantle and remobilized by mantle plumes.

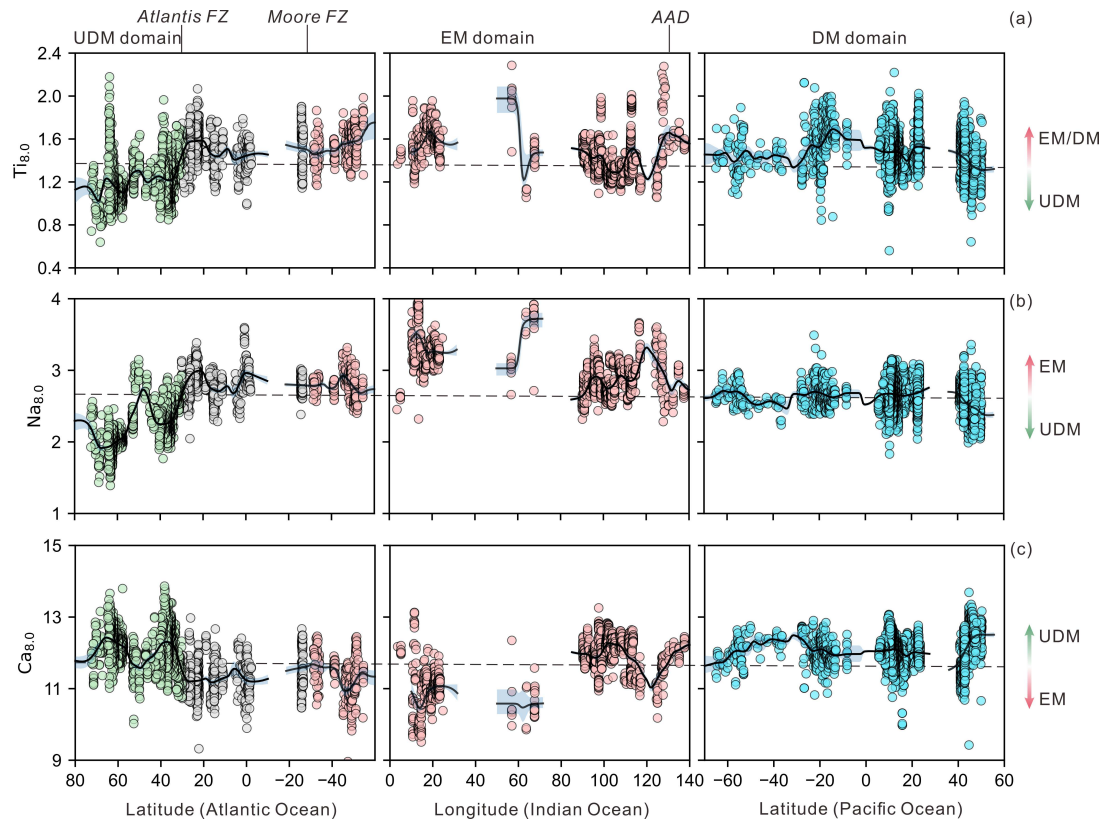


Supplementary Fig. 1 Isotopic compositions of MORBs and near-ridge OIBs/OPBs. a,b, ϵ_{Hf} versus ϵ_{Nd} for OIB/OPB data from ultra-depleted mantle domain (a) and depleted mantle domain (b). **c,d,** $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for OIB/OPB from ultra-depleted mantle domain (c) and depleted mantle domain (d). **e,f,** ϵ_{Hf} versus $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ for OIB/OPB from ultra-depleted mantle domain (e) and depleted mantle domain (f). Isotopic data of the PREMA are from Jackson and Carlson¹⁹ and Giuliani

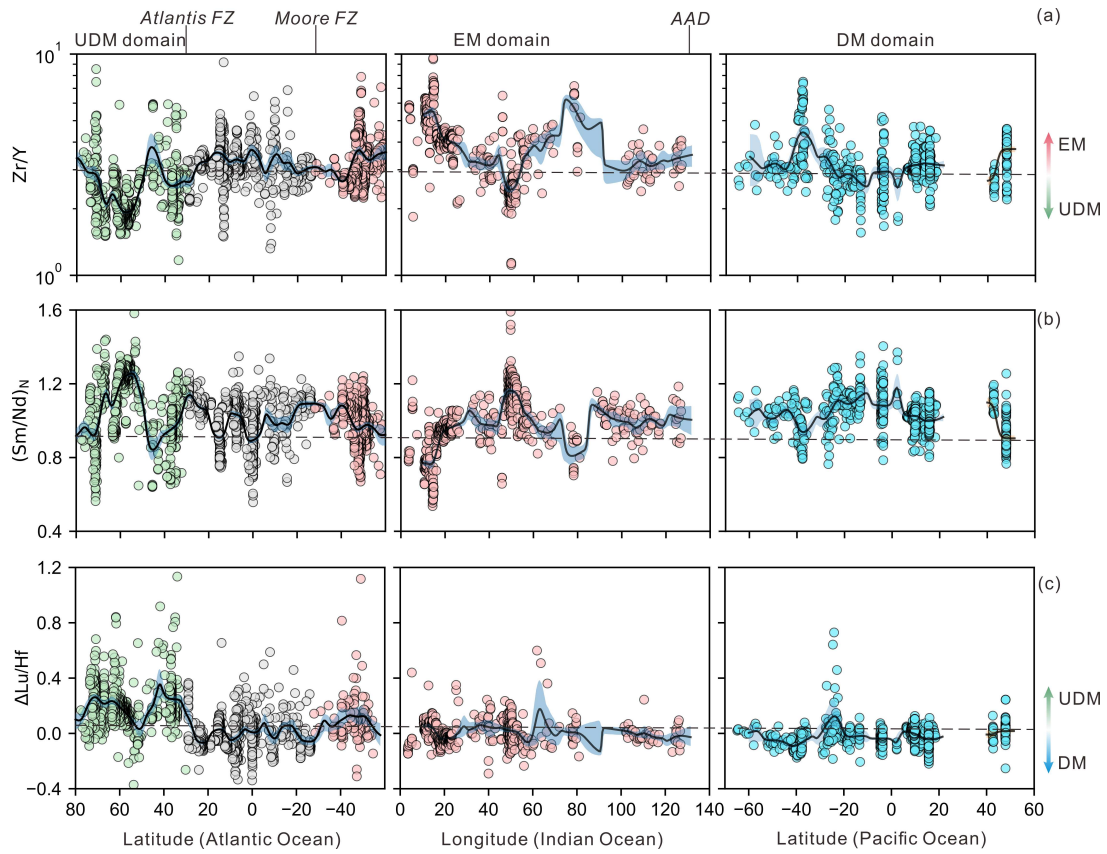
et al.²⁰. Contours denote kernel-density estimates of MORB data from the northern Atlantic Ocean, Pacific Ocean and southern Atlantic and Indian Oceans. Points show the relative density of OIB and OPB data estimated using a Gaussian kernel. Geochemical data of oceanic basalts are listed in [Supplementary Table 1](#).



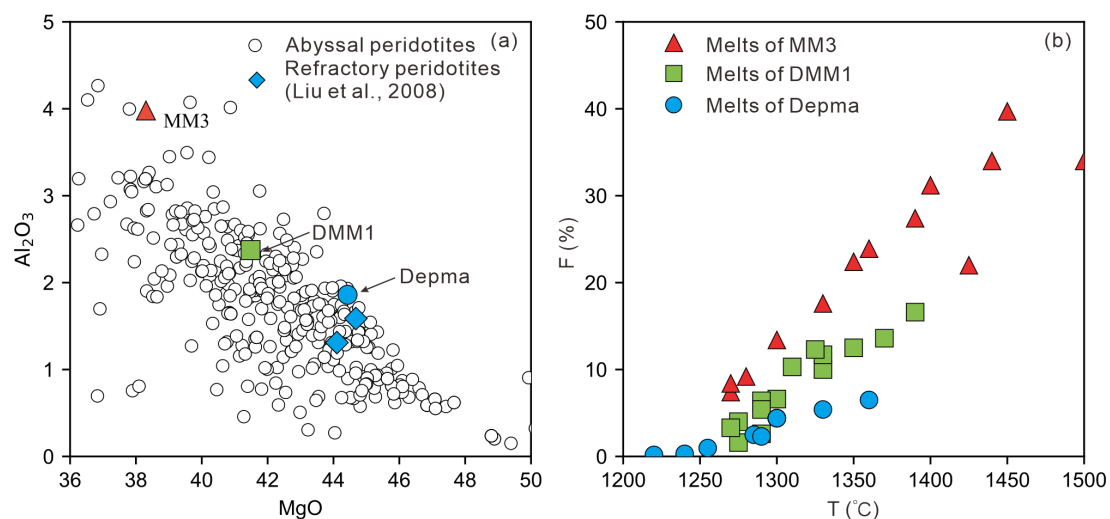
Supplementary Fig. 2 Subduction-related trace-element compositions of MORBs and OIBs/OPBs. a, Ba/Th versus Nb/Nb*. $Nb/Nb^* = Nb_N / (Th_N \times La_N)^{0.5}$. b, Th/U versus Nb/Nb*.



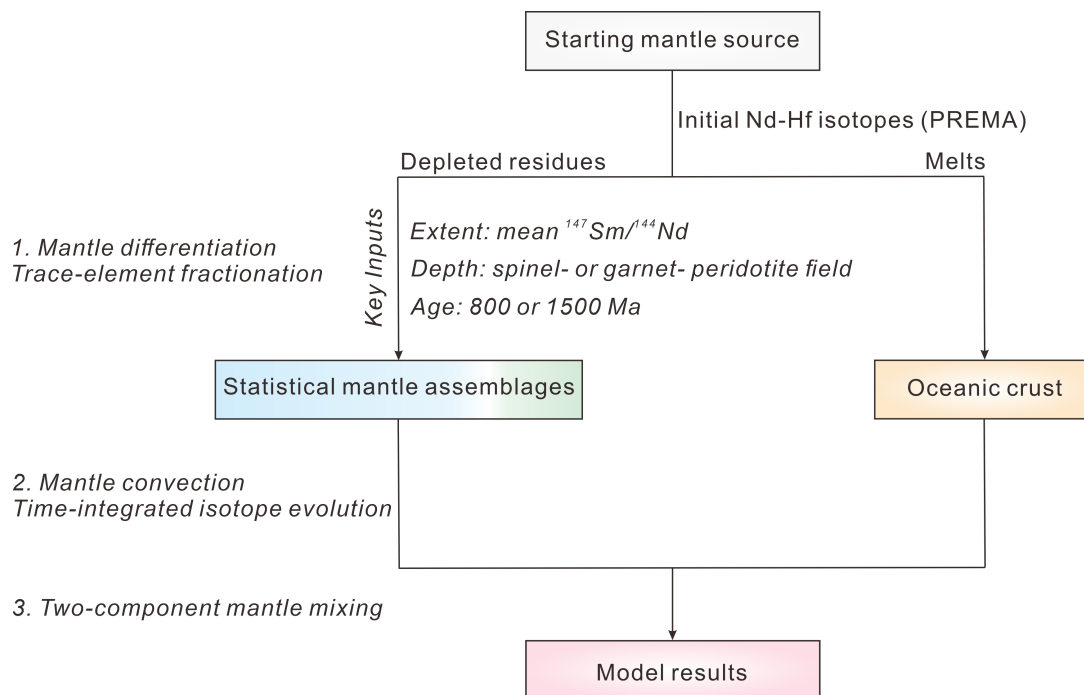
Supplementary Fig. 3 Major-element variations of oceanic basalts along global mid-ocean ridges. a–c, $Ti_{8.0}$, $Na_{8.0}$ and $Ca_{8.0}$, respectively, from Gale et al.⁵⁵. Black curves show Gaussian-kernel regressions of MORB data, and shaded envelopes indicate 95% bootstrap confidence intervals.



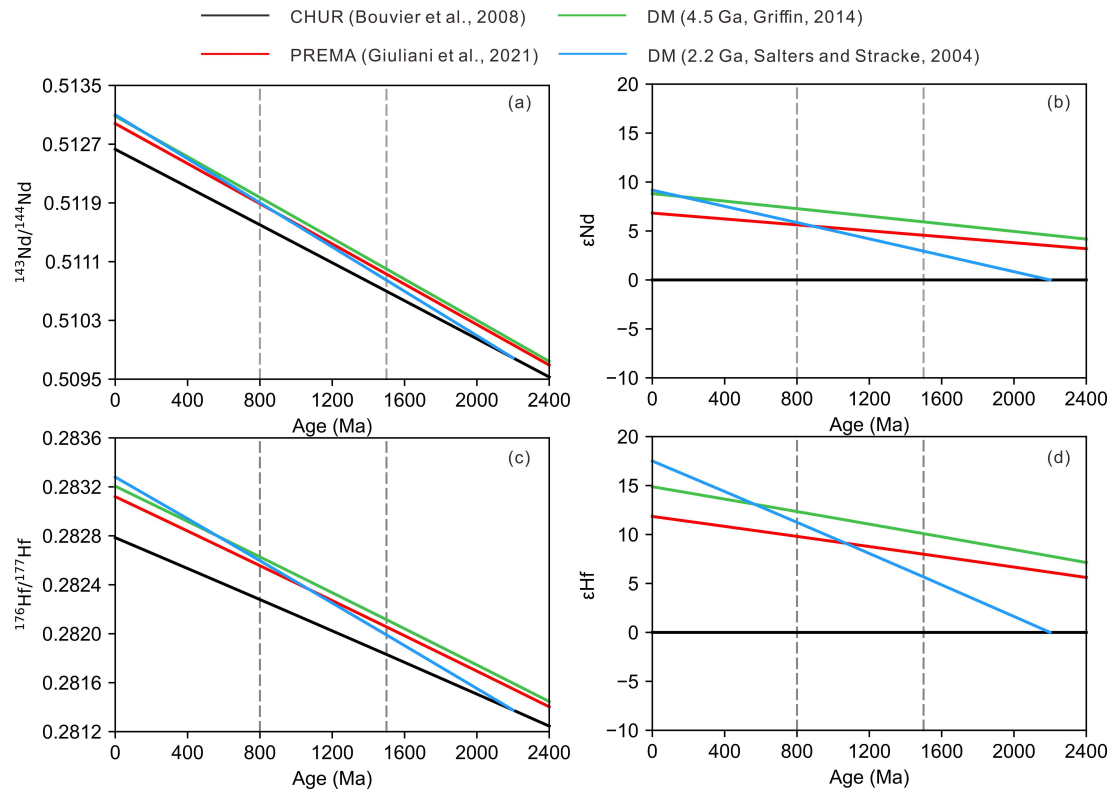
Supplementary Fig. 4 Trace-element variations of oceanic basalts along global mid-ocean ridges. a, Zr/Y . b, $(Sm/Nd)_N$. c, $\Delta Lu/Hf$. $\Delta Lu/Hf = (Lu/Hf)_N - (1.17 \times (Sm/Nd)_N - 0.41)^{56}$. Black curves show Gaussian-kernel regressions of MORB data, and shaded envelopes indicate 95% bootstrap confidence intervals.



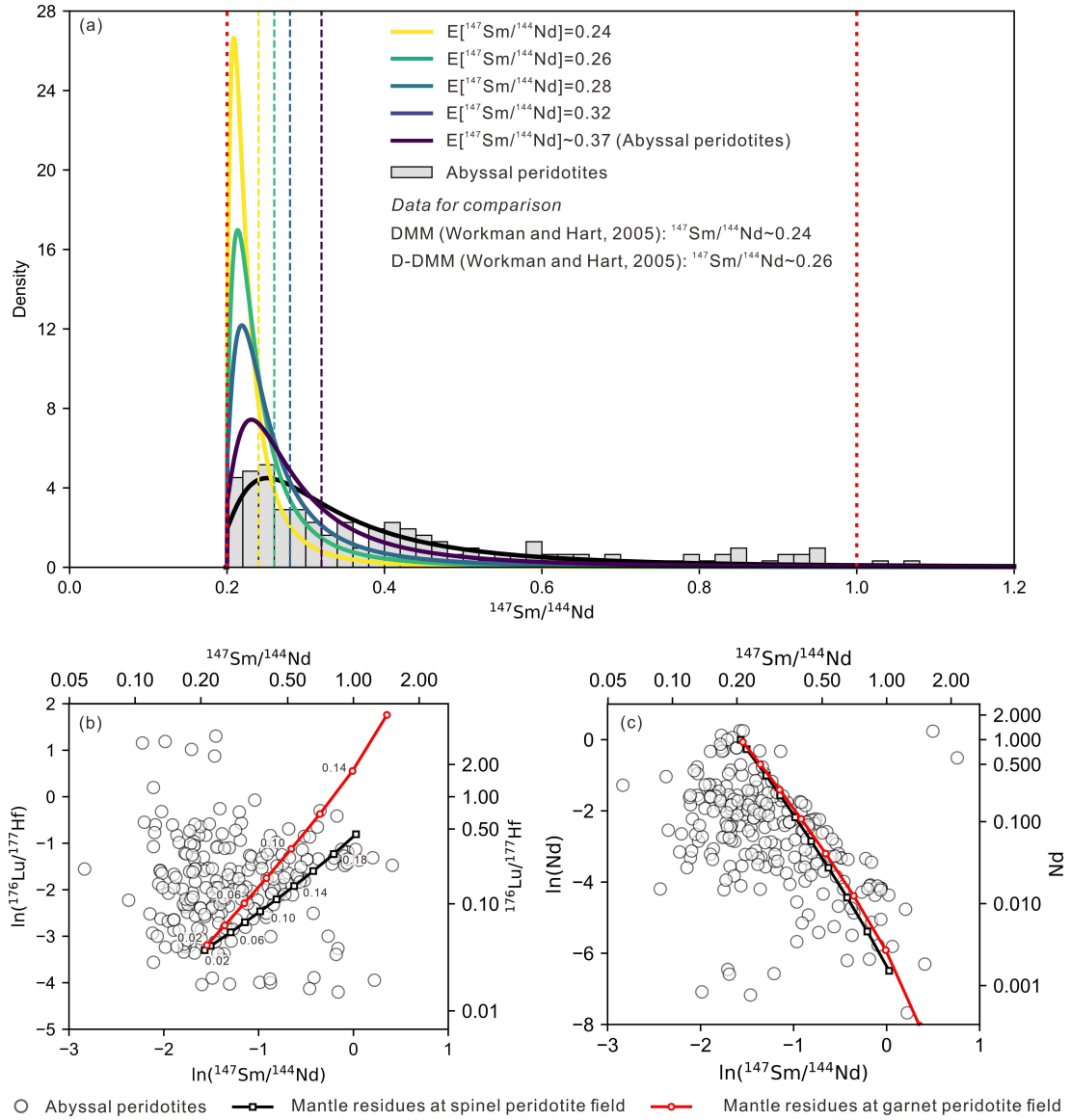
Supplementary Fig. 5 Major-element compositions of abyssal peridotites and experimental mantle melting. a, Major-element compositions of abyssal peridotites. **b,** Experimental melting of fertile and depleted mantle. Abyssal peridotite and experimental data are provided in [Supplementary Tables 2 and 4](#), respectively.



Supplementary Fig. 6 Schematic workflow of the two-component mantle mixing model.



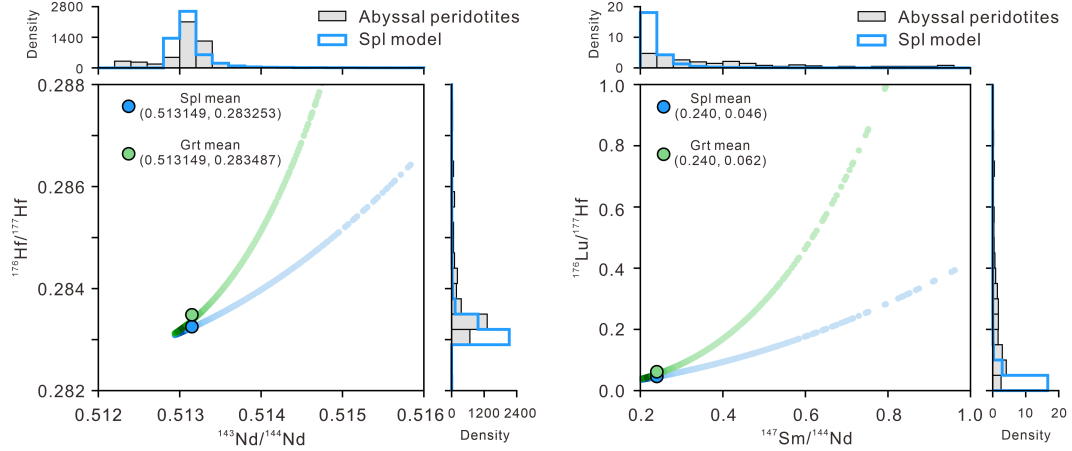
Supplementary Fig. 7 Nd and Hf isotope evolution of distinct mantle reservoirs. a,
 $^{143}\text{Nd}/^{144}\text{Nd}$ versus Age. **b,** ϵ_{Nd} versus Age. **c,** $^{176}\text{Hf}/^{177}\text{Hf}$ versus Age. **d,** ϵ_{Hf} versus
Age. DM, depleted mantle formed at 4.5 Ga⁵² or at 2.2 Ga⁵⁷.



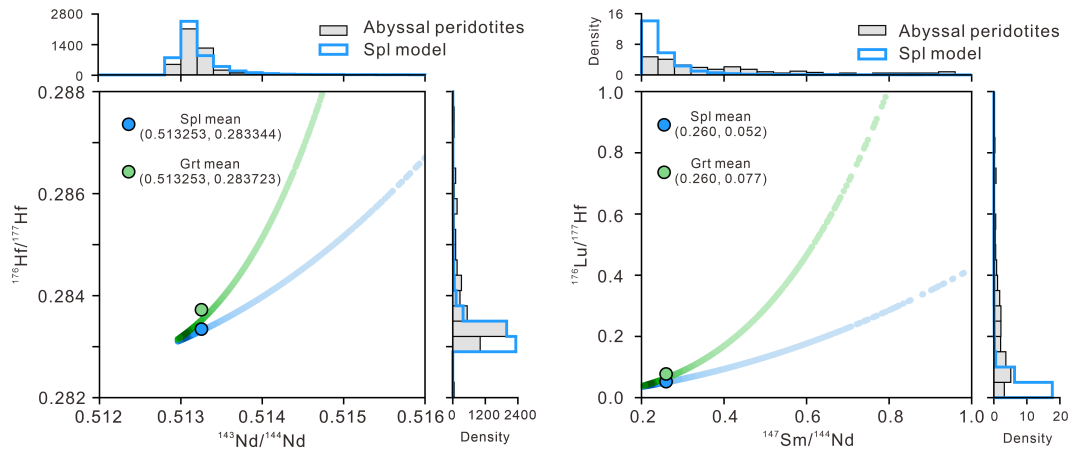
Supplementary Fig. 8 Trace element compositions of abyssal peridotites. a, A histogram for $^{147}\text{Sm}/^{144}\text{Nd}$ of abyssal peridotites under truncation of [0.2, 1.0] with the fitted Johnson SU distribution. **b,** $\ln(^{176}\text{Lu}/^{177}\text{Hf})$ versus $\ln(^{147}\text{Sm}/^{144}\text{Nd})$. **c,** $\ln(\text{Nd})$ versus $\ln(^{147}\text{Sm}/^{144}\text{Nd})$. Black and red curves show residues predicted by dynamic melting models in the spinel- and garnet-peridotite stability fields, respectively. The models use a primitive-mantle starting composition from McDonough and Sun²¹. Abyssal peridotite data are provided in [Supplementary Table 2](#).

Mantle depletion from a PREMA source

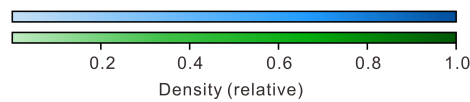
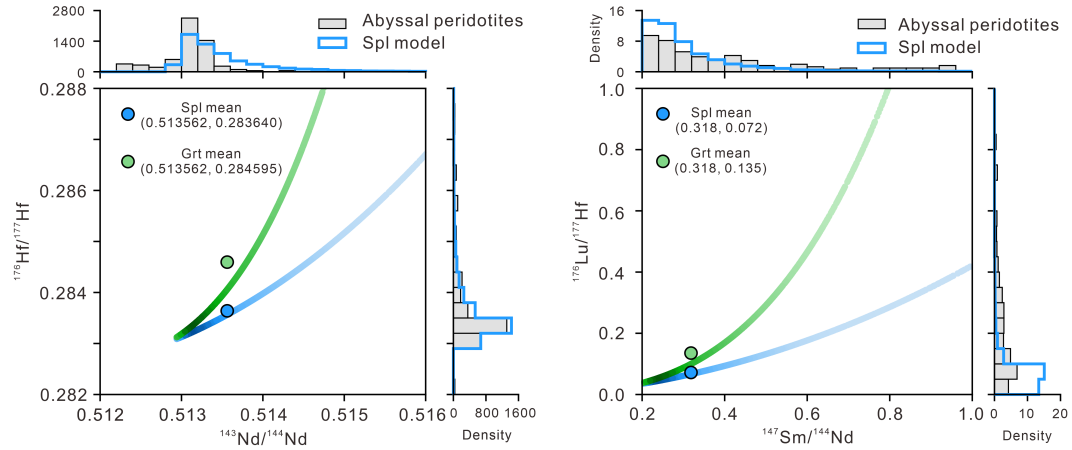
(a) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.24$; Age: 800 Ma



(b) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.26$; Age: 800 Ma



(c) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.32$; Age: 800 Ma

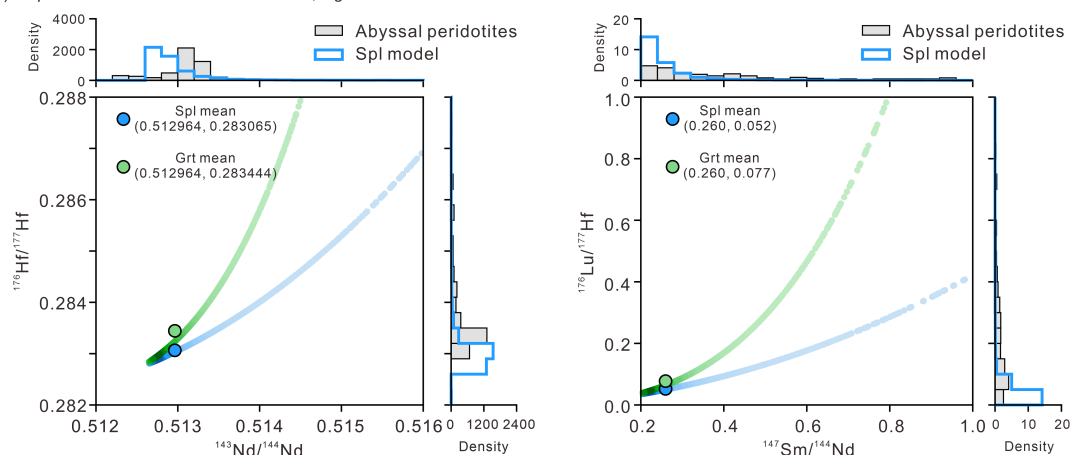


Supplementary Fig. 9 Modelled isotopic and trace-element compositions of mantle assemblages derived from a PREMA-like reservoir at 800 Ma. a–c, Monte Carlo simulations for mantle assemblages with mean depletion extents of $^{147}\text{Sm}/^{144}\text{Nd} = 0.24$,

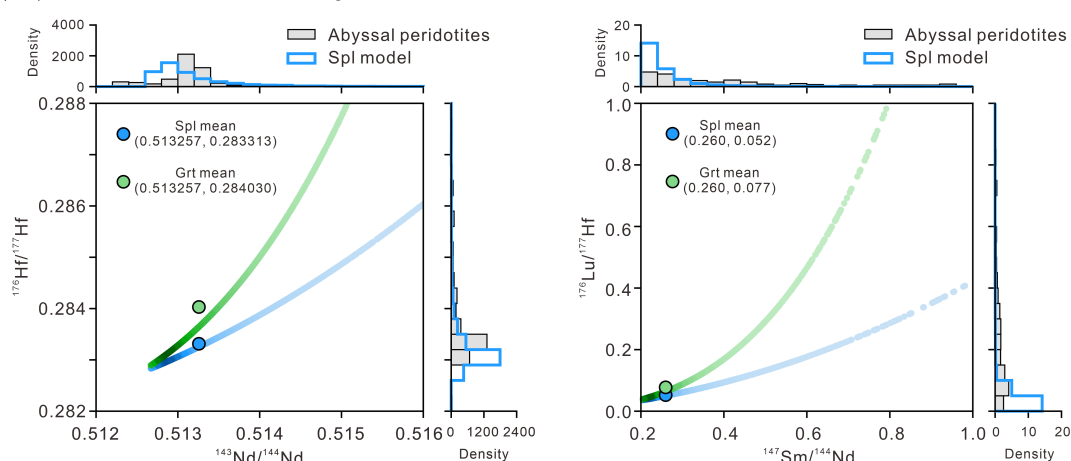
0.26, and 0.32, respectively. Each assemblage was constructed using a distribution parameterized from abyssal peridotite data ([Supplementary Fig. 8](#)). Blue and green points show simulated compositions of residues formed in the spinel- and garnet-peridotite stability fields, respectively, where color intensity indicates relative point density. Filled circles denote the mean compositions of the simulated spinel- and garnet-field residues. Histograms along the top and right margins compare simulated spinel-field residues with observed abyssal peridotites.

Mantle depletion from a CHUR-like source

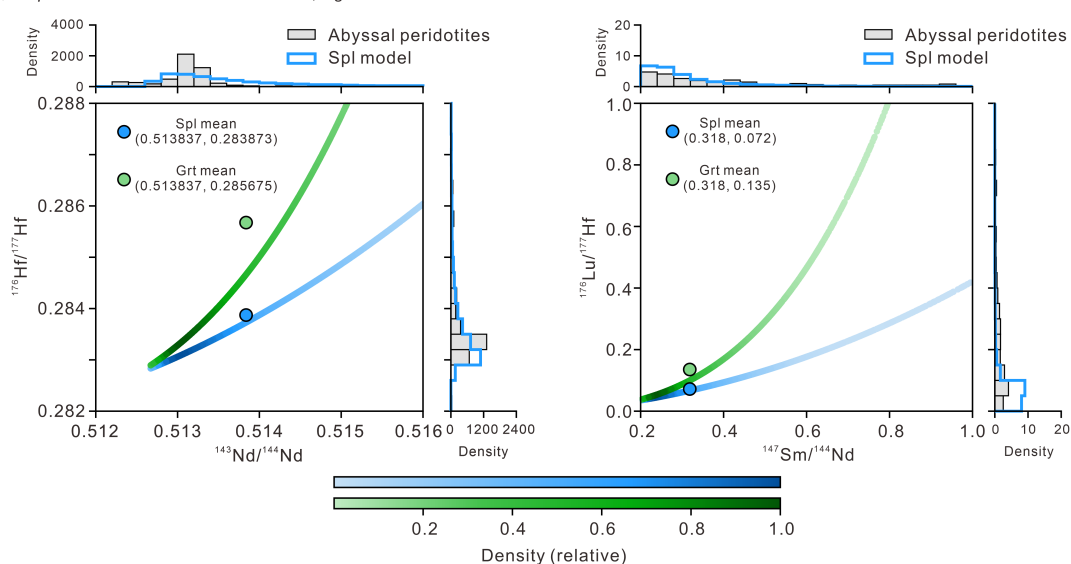
(a) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.26$; Age: 800 Ma



(b) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.26$; Age: 1500 Ma



(c) Depletion extent: $^{147}\text{Sm}/^{144}\text{Nd} = 0.32$; Age: 1500 Ma

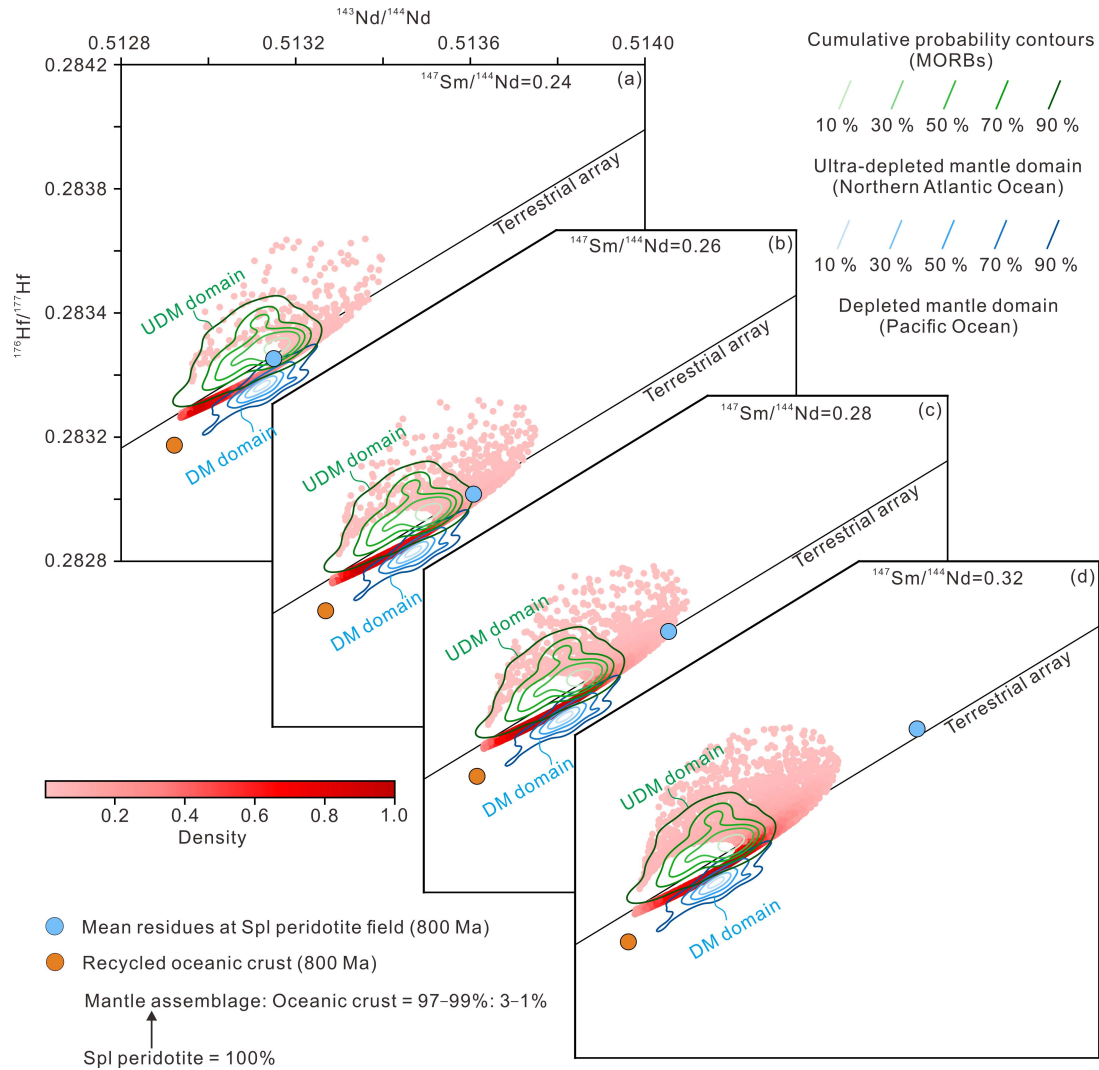


567

568 **Supplementary Fig. 10 Modelled isotopic and trace-element compositions of**
569 **mantle assemblages derived from a CHUR-like reservoir at 800 and 1500 Ma. a–**
570 **b, Monte Carlo simulations for mantle assemblages with mean depletion extents of**

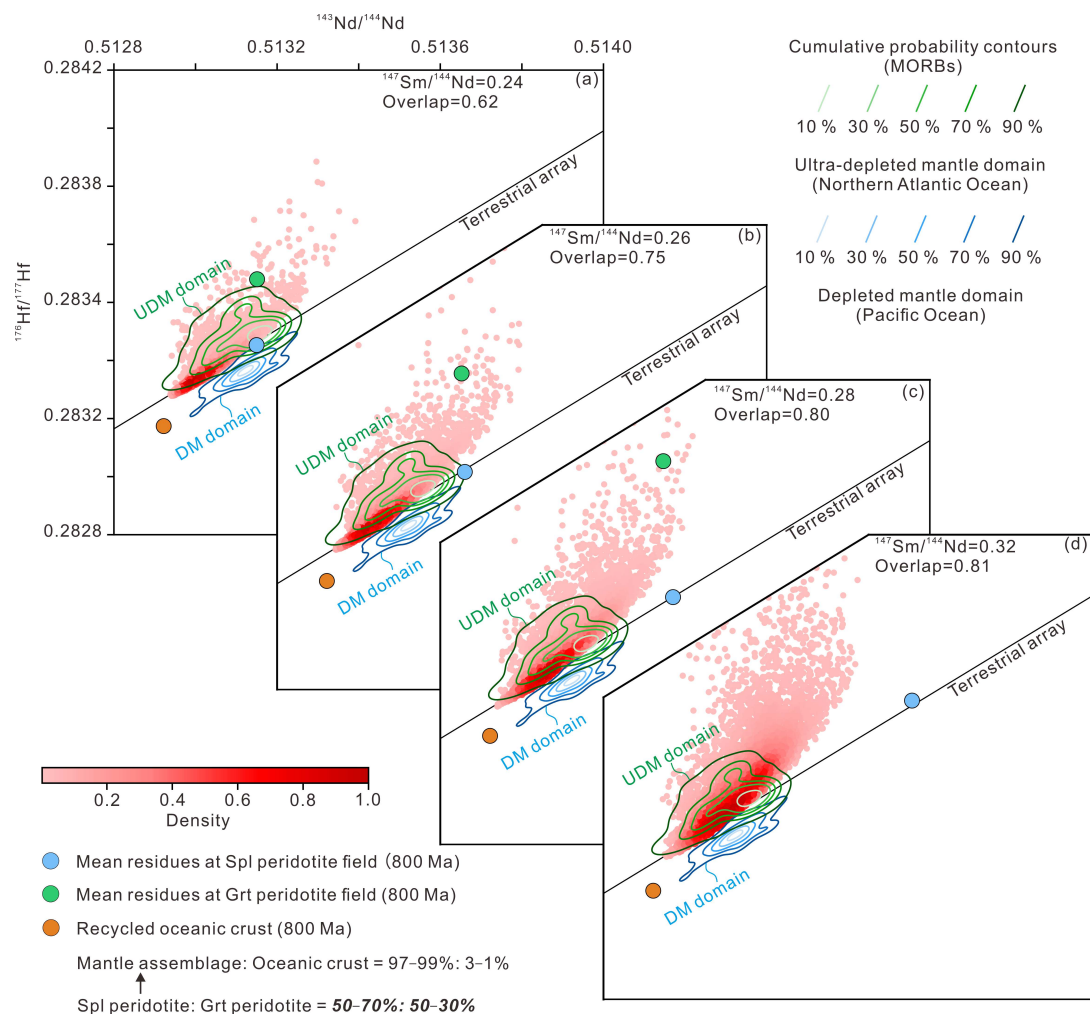
571 $^{147}\text{Sm}/^{144}\text{Nd} = 0.26$ formed at 800 and 1500 Ma. **c**, Monte Carlo simulations for a mantle
572 assemblage with mean depletion extents of $^{147}\text{Sm}/^{144}\text{Nd} = 0.32$ formed at 1500 Ma.
573 Each assemblage was constructed using a distribution parameterized from abyssal
574 peridotite data ([Supplementary Fig. 8](#)). Blue and green points show simulated
575 compositions of residues formed in the spinel- and garnet-peridotite stability fields,
576 respectively, where color intensity indicates relative point density. Filled circles denote
577 the mean compositions of the simulated spinel- and garnet-peridotite field residues.
578 Histograms along the top and right margins compare simulated spinel-field residues
579 with observed abyssal peridotites.

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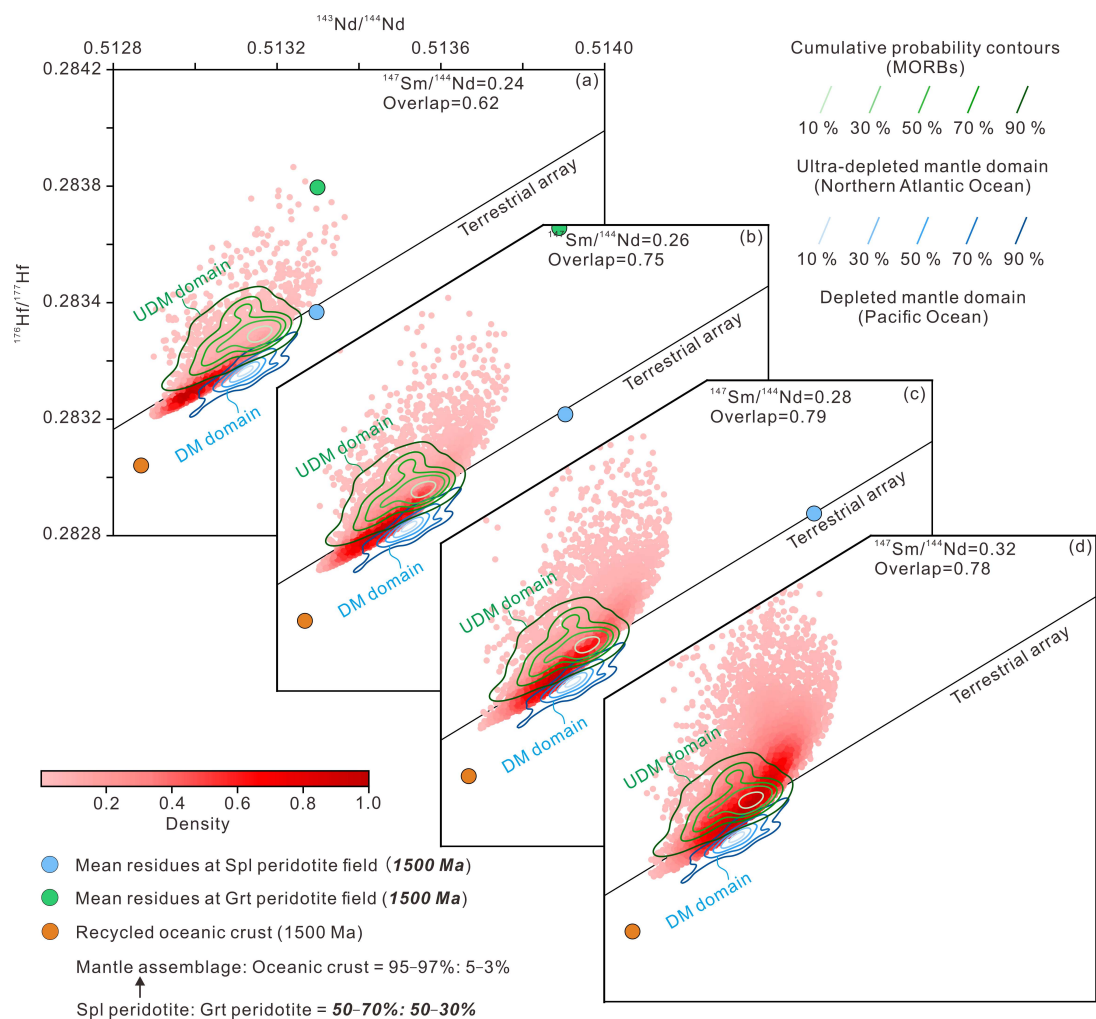
582

583 **Supplementary Fig. 11 Modelled isotope compositions of mixtures of spinel-**
 584 **peridotite field mantle residues and recycled oceanic crust. a–d, Monte Carlo model**
 585 **results of mantle assemblages with mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.24, 0.26, 0.28 and 0.32,**
 586 **respectively (Supplementary Fig. 8). Mantle assemblages formed in pure spinel**
 587 **peridotite field at 800 Ma are mixed with a minor recycled oceanic crust component**
 588 **(1–3%) of the same age. Modelled results are shown as density-colored scatters, where**
 589 **color intensity reflects relative point density. Blue circles indicate mean isotopic**
 590 **compositions of spinel-peridotite field residues, respectively, whereas orange circles**
 591 **denote recycled oceanic crust. Blue and green contours represent the observed DM and**
 592 **UDM domains, respectively. The black line represents the terrestrial array¹⁷.**



Supplementary Fig. 12 Modelled isotope compositions of mantle mixtures containing 30–50% garnet-peridotite field residues in the mantle component. a–d, Monte Carlo model results of mantle assemblages with mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.24, 0.26, 0.28 and 0.32, respectively (Supplementary Fig. 8). Mantle assemblages formed at 800 Ma comprise variable proportions of spinel- and garnet-peridotite residues (spinel peridotite: garnet peridotite = 70–50%:30–50%), and are mixed with a minor recycled oceanic crust component (1–3%) of the same age. Modelled results are shown as density-colored scatters, where color intensity reflects relative point density. Reported overlap values quantify the areal overlap between the modelled distributions and the observed ultra-depleted mantle (UDM) domain field. Blue and green circles indicate mean isotope compositions of spinel- and garnet-peridotite field residues, respectively; orange circles indicate the recycled oceanic crust component. Blue and green contours represent the observed DM and UDM domains, respectively. The black

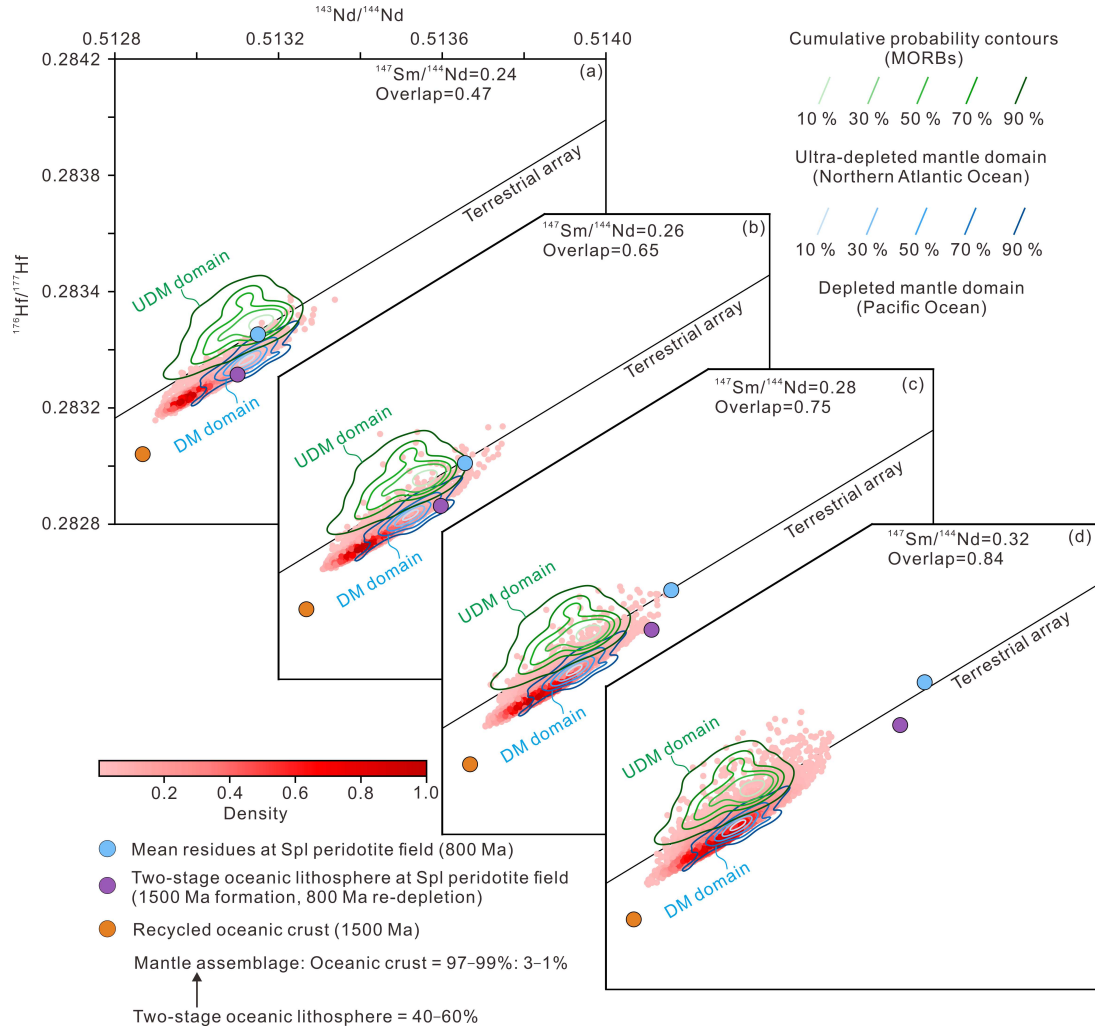
607 line represents the terrestrial array¹⁷.



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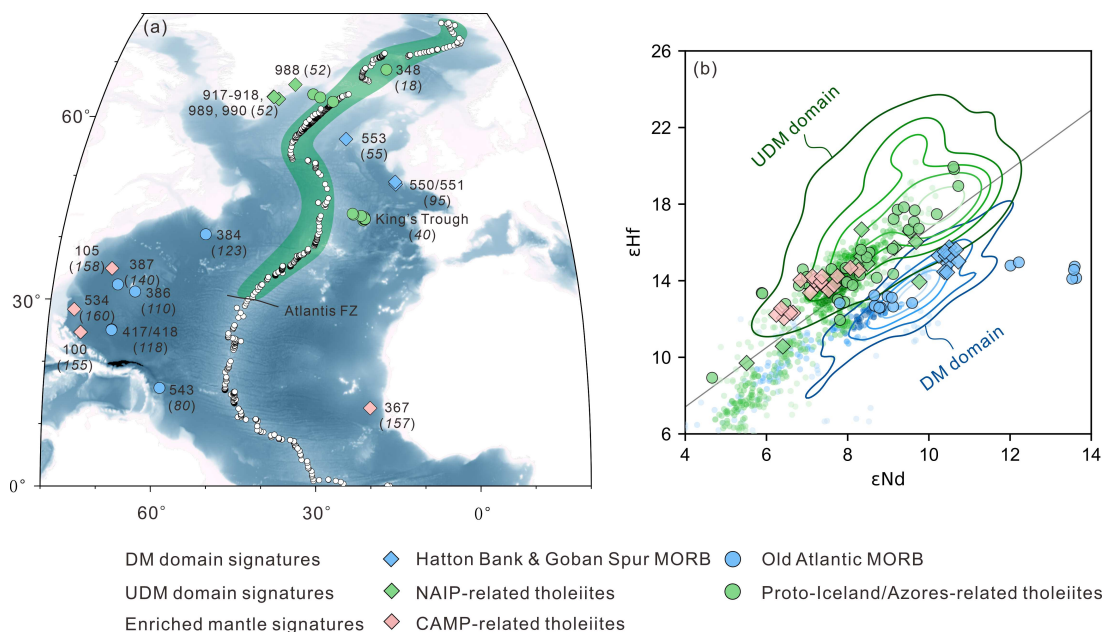
609 **Supplementary Fig. 13 Modelled isotope compositions of mixtures of mantle**
 610 **residues and recycled oceanic crust formed at 1,500 Ma. a–d, Monte Carlo model**
 611 results of mantle assemblages with mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.24, 0.26, 0.28 and 0.32,
 612 respectively (Supplementary Fig. 8). Mantle assemblages formed at 1500 Ma comprise
 613 variable proportions of spinel- and garnet-peridotite residues (spinel peridotite: garnet
 614 peridotite = 70–50%: 30–50%), and are mixed with a minor recycled oceanic crust
 615 component (3–5%) of the same age. Modelled results are shown as density-colored
 616 scatters, where color intensity reflects relative point density. Reported overlap values
 617 quantify the areal overlap between the modelled distributions and the observed ultra-
 618 depleted mantle (UDM) domain field. Blue and green circles indicate mean isotopic
 619 compositions of spinel- and garnet-peridotite field residues, respectively; orange circles

indicate the recycled oceanic crust component. Blue and green contours represent the observed DM and UDM domains, respectively. The black line represents the terrestrial array¹⁷.



Supplementary Fig. 14 Modelled isotope compositions of mantle mixtures incorporating recycled lithospheric residues with two-stage depletion histories. a–d, Monte Carlo model results of mantle assemblages with mean $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.24, 0.26, 0.28 and 0.32, respectively (Supplementary Fig. 8). Mantle assemblages comprise variable proportions of spinel peridotite residues (40–60 %) and recycled oceanic lithosphere (1.5 Ga, 60–40 %) with subsequent melt extraction at 800 Ma, and are mixed with a minor recycled oceanic crust component (1–3 %) of 1.5 Ga. Blue and purple circles show mean isotopic compositions of spinel peridotite residues and two-stage recycled oceanic lithosphere, respectively; orange circles show the recycled

oceanic crust component. Reported overlap values quantify the areal overlap between the modelled distributions and the observed depleted mantle (DM) domain field. Blue and green contours represent the observed DM and UDM domains, respectively. The black line represents the terrestrial array¹⁷.



Supplementary Fig. 15 Distribution and isotopic compositions of tholeiites from the early northern Atlantic Ocean. CAMP, Central Atlantic Magmatic Province; NAIP, North Atlantic Igneous Province.

Supplementary Files

This is a list of supplementary files associated with this preprint. Click to download.

- [SupplementaryTable4.xlsx](#)
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- [SupplementaryTable2.xlsx](#)