**Supplementary Information**

**1. Global oceanic basalt U-series isotope database**

The global database provided in Table S1 compiles published mass spectrometry data for 238U-230Th-226Ra-210Pb isotopes and, where readily available, accompanying radiogenic isotope studies. Trace element ratios of common interest are also included, although additional measurements for the same samples from unrelated studies may have been omitted, and a full major and trace element data set has not been included as it is beyond the scope of this review. The database has been further synthesized such that each row represents a single sample, even where multiple analyses have been published across several studies. Overall, there are few replicates across multiple studies for a single sample, but in a number of cases replicate measurements for U-series isotopes have been published within a single published article; in these cases, an averaged value is presented. Users are cautioned that where multiple references are listed for a single sample, the original literature should be consulted to guarantee that the proper references are fully credited for all measurements.

Where possible, accompanying location, sublocation, on- or off-axis position (for mid-ocean ridge basalts), and sampling depth (for submarine samples) have been incorporated for easy reference. Age data are provided based on the maximum possible or known eruption age reported for each sample, based on historic eruption records, stratigraphic constraints, or U-series disequilibria. Where historic eruption dates are known, the age of the sample is calculated for the calendar year when the file was last opened, and is shown in the “Max age (yrs BP)” column.

Four additional criterion-based data columns are included in the database. Three of these use the maximum reported age of a sample to determine whether existing age constraints are sufficient for a U-series nuclide pair to be considered unaffected by radioactive decay since eruption, using a limiting constraint set by the user and expressed in terms of the half-lives of the nuclides being considered. These columns, “Th age constraints,” “Ra age constraints,” and “Pa age constraints,” return a simple Yes/No response for whether such an age constraint exists (Yes) or not (No). Users can impose a number of half-lives within which the age of a sample is considered well-constrained in the “Age Constraints” worksheet; the default value is one half-life of the daughter nuclide.

The fourth such column, “Unaltered?” queries whether (234U/238U) activity ratios have been measured, and if so, whether they are within uncertainty of equilibrium (± 1%). Uranium isotope disequilibrium occurs in subaerial basalts due to continental weathering, and in submarine rocks due to chemical interaction with seawater or seawater-derived materials; i.e., it is a sensitive indicator of weathering and alteration processes (e.g., Aumento, 1971; Bourdon et al., 2000; Cheng et al., 2000; Henderson et al., 1993; Ku et al., 1977; Macdougall et al., 1979; Porcelli and Swarzenski, 2003; Robinson et al., 2004; Sims et al., 2003; Suhr et al., 2018; Thurber, 1962; Turekian and Bertine, 1971). The age- and alteration filtering methods used in these columns are the same as those used to produce the diagrams of filtered global MORB and OIB data shown in Figures 3 and 4.

Supplementary Figures S3-S12 show the complete (230Th/238U), (226Ra/230Th), and (231Pa/235U) melt modeling results for the model calculations referenced in Section 5.2.5, including both thermal isolation and thermal equilibration between two coexisting lithologies, dynamic and RPF melting of peridotite and pyroxenite lithologies, two different pyroxenite compositions (i.e., recycled gabbro and Si-poor, olivine-bearing pyroxenite), mantle potential temperatures of 1300ºC and 1550ºC, lithospheric lids corresponding to final pressures of melting of 0.5 and 2.0 GPa, and a range of solid mantle upwelling rates and maximum residual melt porosities; this last set of variables is indicated with different color lines within each figure panel. Fields indicate the compositions of measured MORB and OIB for comparison, as shown and referenced in Figures 3 and 4.

**2. Summary of magma generation and transport effects on U-series isotopes in a homogeneous peridotite source**

In the dynamic model of McKenzie (1985a), melting takes place during decompression in a vertical section of upwelling mantle (Figure 1a) and produce the geochemical effects expected in near-fractional or “continuous” melting, whereby incompatible elements (including the U-series nuclides) are rapidly and efficiently stripped out of the solid residue and concentrated in the earliest melt fractions produced (e.g., McKenzie, 1985; Williams and Gill, 1989). As described in the main text, Uranium-series isotopic disequilibria arise when parent and daughter nuclides have different residence times in the melting column. During the prolonged residence time of U in the presence of residual garnet, additional 230Th is continuously produced by radioactive ingrowth due to local disequilibrium in the residue, generating far larger 238U-230Th disequilibrium over time than instantaneous fractional melting (e.g., Williams and Gill, 1989). The extent of U-Th disequilibria is thus positively correlated with the garnet mode and the overall melting time, a product of solid mantle upwelling rate, the length of the melting column, and the initial (solidus) depth of melting.

226Ra has a half-life of 1600 years, so any 230Th-226Ra disequilibrium produced by melting is only detectable if erupted and measured within 8000 years (Figure S1). Because dynamic melting models predict that most disequilibria are produced in the initial stages of melting, elevated (226Ra/230Th) ratios like those observed in most oceanic basalts (see below) must survive transport from the deep melting column all the way to the surface. For 230Th-226Ra disequilibria and deep melting regimes, that survival requires particularly rapid melt transport, from 10s of meters to several kilometers per year (e.g., Stracke and Bourdon, 2009; Stracke et al., 2006). For comparison, the maximal solid mantle upwelling rate calculated for Hawaii (i.e., at the plume axis) is only 9.6 m/yr. (Hauri et al., 1994), and expected solid mantle upwelling beneath slow-spreading ridges can be as low as 10 mm/year (e.g., Kelemen et al., 1997). While magma can migrate appreciably faster than the upwelling solid matrix, the particularly high liquid ascent rates required by dynamic melting and 230Th-226Ra disequilibria are probably unlikely in some circumstances. Channelized transport presents a possible solution to this restriction: dynamic melting could potentially occur at great depths in the mantle and still preserve disequilibria as large as those observed in global marine basalts (Figures 3-4), if channels permitting rapid transport are present early and deep enough in the melting process (e.g., Hart, 1993). Dunite bodies found in ophiolites have been described as potential “fossils” of such channels (e.g., Kelemen et al., 1997). Early petrological modelling in a homogeneous peridotitic source suggested that dunite channels could only form at relatively low pressure (0.5 GPa for a solidus at 1.9 GPa, about 0.8 GPa for a solidus at 3.5 GPa) (Asimow and Stolper, 1999). However, numerical modeling of reactive infiltration instabilities support the possibility of channels deep in the melt regime (e.g., Jull et al., 2002; Liang, 2008; Spiegelman et al., 2001), and extreme trace element depletions in abyssal peridotites as well as major element systematics in basalts have been cited as support for a melt regime dominated by near-fractional processes in the mantle overall (e.g., Hart and Dunn, 1993; Johnson et al., 1990; Langmuir et al., 1992; Sobolev and Shimizu, 1993). Moreover, the presence of lithological heterogeneity or volatiles in the source can favor an early development of high permeability channels (Keller et al., 2017; Weatherley and Katz, 2012, 2016).

As described in the main text, the alternative “reactive porous flow” (RPF) model suggested by Spiegelman and Elliott’s (1993) model alleviate the most extreme transport rate requirements imposed by dynamic melting models and the preservation of large 230Th-226Ra disequilibrium in partial melts generated in the mantle (e.g., Elkins et al., 2014; Sims et al., 2002). Spiegelman and Kelemen (2003) further calculated extremely diverse trace element makeup of individual instantaneous melt droplets along capillary paths in the mantle, suggesting that prior to homogenization by mixing, partial melt compositions could be extremely diverse due to transport variations. Recent studies modeling melt transport in relatively homogeneous mantle also suggest that a homogeneous source would likely be dominated by RPF processes, rather than channelized flow (Katz and Weatherley, 2012; Weatherley and Katz, 2012, 2016), so this model may be particularly relevant to more homogeneous (e.g., purely peridotitic or well-mixed) mantle regimes, like some mid-ocean ridge settings.

While the extreme trace element depletion of abyssal peridotites suggests a near-fractional mantle melting regime (e.g., Johnson et al., 1990), trace element depletion can be reconciled with evidence for porous flow scenarios if both occur simultaneously in a single magma transport regime (i.e., two mantle flow regimes with different porosities, effectively behaving as low-porosity porous flow and high-porosity channelized zones) (e.g., Kelemen et al., 1997; Kelemen et al., 1995; Liang, 2008; Lundstrom, 2000; Sims et al., 2002; Spiegelman et al., 2001). Jull et al. (2002) estimated that as much as 50% of magma may be transported by RPF-dominated, low-porosity zones adjacent to high-porosity channels, and, owing to the effect of thermal diffusion, Weatherley and Katz (2016) observed in their model calculations that melts in low-porosity regions could largely remain segregated from melts transported by channels, similar to Spiegelman and Kenyon’s (1992) predictions.

Lundstrom et al. (2000) and Lundstrom (2000) argued that chemical interactions during porous flow magma transport through the uppermost mantle could produce the observed trace element patterns in residual mantle rocks (in agreement with Warren, 2016, and references therein). Niu (1997) likewise identified evidence for grain-scale porous flow in mantle-derived rocks. Furthermore, global MORB have been observed to lie along a negative correlation in (226Ra/230Th) vs. (230Th/238U) diagrams (Lundstrom et al., 1999; Sims et al., 2002). Sims et al. (2002) confirmed this correlation for age-constrained samples from the East Pacific Rise (EPR) that have relatively constant radiogenic isotope compositions indicative of a homogeneous mantle source; the correlation has also been observed in MORB from the Juan de Fuca-Gorda and the eastern Gakkel Ridge, among others, suggesting a global pattern for ridge basalts that are unaffected by proximity to hotspots (e.g., Batiza and Niu, 1992; Elkins et al., 2014; Goldstein et al., 1989; Kelemen et al., 1997; Lundstrom et al., 1999; Volpe and Goldstein, 1993). Sims et al. (2002) and Jull et al. (2002) proposed that the negative correlation is best explained by a two-porosity transport regime, where melts produced by peridotite melting at great depths, in the presence of residual garnet, have high (230Th/238U) but have lost initial 226Ra excesses to radioactive decay during transport in high-porosity channels; while melts that travel slowly via RPF have low (230Th/238U) and high (226Ra/230Th), recording continued shallow melting in the absence of garnet by porous flow. Such a scenario is mechanistically consistent with fluid transport and reactive infiltration predictions (e.g., Kelemen et al., 1997; Spiegelman et al., 2001), although melt flow from the porous matrix to the melt channel has been difficult to prescribe *a priori*.

In contrast, several studies have explored the possibility that much of the observed variation in basalt composition could be explained by melting without achieving full chemical equilibrium (e.g., Bourdon and Van Orman, 2009; Spiegelman and Kenyon, 1992; Van Orman et al., 2002, 2006). Melt-rock interactions likely play a role in producing the geochemical makeup of oceanic basalts (e.g., Brunelli et al., 2014; Elkins et al., 2014; Elliott and Spiegelman, 2003; Godard et al., 2008; Lambart et al., 2012; Lissenberg and Dick, 2008; Seyler et al., 2007; Warren et al., 2009). Chemical disequilibrium and diffusion models (Bourdon and Van Orman, 2009; Saal and Van Orman, 2004; Van Orman and Saal, 2009; Van Orman et al., 2006) explicitly consider the effects of chemical fractionation due to slow diffusion between partial melts and the surrounding solid matrix as a mechanism for producing U-series isotopic disequilibrium, within both the mantle source region and the overlying lithosphere and oceanic crust. They argue that under the right circumstances, the interactions between partial melts of mantle peridotites, and oceanic crustal cumulates containing pyroxene and feldspar grains of various ages, could explain much of the observed U-series isotope disequilibrium data for basalts, including the (226Ra/230Th) vs. (230Th/238U) negative array for MORB. Cooper et al. (2003), Sims et al. (2003), Turner et al. (2012), and Elkins et al. (2014), however, observed that concentrations of Ra in crustal cumulate minerals are insufficient to explain all of the observed data by diffusion and chemical disequilibrium processes alone. While chemical disequilibrium due to melt-rock interactions is likely an important process in basalt genesis, its effect on U-series isotope disequilibria is expected to be small for most magmas due to the low crustal cumulate concentrations, and producing the MORB array would require a highly coincidental alignment of factors for most oceanic basalts (Van Orman et al., 2006), unlikely to occur in a global, systematic way. We thus suggest that while chemical diffusion effects may be important for some locations or basalts, on the whole diffusion does not *systematically* control U-series isotopes in global oceanic basalts.

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