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The amount of recycled crust in sources of mantle-derived melts

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# The Amount of Recycled Crust in Sources of Mantle-Derived Melts

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Plate tectonic processes introduce basaltic crust (as eclogite) into the peridotitic mantle. The proportions of these two sources in mantle melts are poorly understood. Silica-rich melts formed from eclogite react with peridotite, converting it to olivine-free pyroxenite. Partial melts of this hybrid pyroxenite are higher in nickel and silicon but poorer in manganese, calcium, and magnesium than melts of peridotite. Olivine phenocrysts' compositions record these differences and were used to quantify the contributions of pyroxenite-derived melts in mid-ocean ridge basalts (10 to 30%), ocean island and continental basalts (many >60%), and komatiites (20 to 30%). These results imply involvement of 2 to 20% (up to 28%) of recycled crust in mantle melting.

It is widely accepted that the heterogeneity of the convecting mantle observed in the composition of mantle-derived magmas is largely due to subduction and recycling of oceanic crust into the deep mantle (1, 2). To understand the role of crustal material in creating compositional heterogeneities in the mantle and to evaluate the geodynamical consequences of this contribution, one must quantify the crustal

input to the mantle sources of common, mantle-derived magmas in mid-oceanic ridges basalts (MORBs), ocean islands (OIBs), and large igneous provinces (LIPs). It is not possible to use incompatible element abundances in basalts to constrain the proportion of recycled component in the magma source because concentrations of these elements are also sensitive to the extent of melting. Similarly, the use of isotope ratios for making such quantitative estimates is compromised by the isotopic variability of subducted materials involved in the recycling process (2). We used an alternative approach based on a combination of major elements and compatible trace elements in parental melts, because these are more uniform in the mantle and are strongly controlled by the residual phases in equilibrium with partial melts (3–5).

Our method has its basis in the experimental and theoretical prediction that high-pressure ( $P > 3.0$  GPa) melting of typical recycled oceanic crust (in the form of eclogite with a separate  $\text{SiO}_2$  phase) and reaction of this melt with peridotite produces olivine-free pyroxenite (5). We show that further melting of this hybrid lithology in the absence of residual olivine is more voluminous than the melting of peridotite (at a given pressure and temperature) and that pyroxenite-derived melts are characteristically enriched in Si and Ni but depleted in Mg, Ca, and Mn compared with their peridotite-derived counterparts. This difference arises because olivine principally controls the composition of melt produced in peridotite, whereas pyroxene mainly controls the composition of melt from olivine-free hybrid pyroxenite (5–8). Experimental data predict (9) that, as such pyroxenite-derived melts rise toward the surface, the decrease in pressure causes their saturation in olivine. This olivine is unusually Ni rich and Mn and Ca poor. With use of a new, large data set of high-precision analyses of

olivine phenocrysts from OIBs, LIPs, MORBs, and komatiites, we show that hybrid pyroxenite is a common lithology in upwelling mantle and a major contributor to tholeiitic (silica-saturated) and transitional (moderately silica-undersaturated) magmas of OIBs and LIPs emplaced on thick oceanic or continental lithosphere.

**Olivine data set.** We use olivine phenocrysts as probes of parental melt composition, because olivine is the first phase to precipitate at low pressures in almost all mantle-derived magmas and because its forsterite content is an excellent measure of the degree of fractional crystallization allowing reconstruction of the parental melt composition.

Olivine phenocrysts were analyzed by electron microprobe using high probe currents and long counting times (10). This procedure routinely yields detection limits of around 6 to 15 parts per million (ppm) and errors (2 standard errors) of 15 to 30 ppm for trace elements (Ni, Ca, Mn, Cr, Co, and Al) and 0.01 mole percent (mol %) for forsterite content [defined as  $Fo = \text{Mg}/(\text{Mg} + \text{Fe})$ ], checked by repeated analysis of San Carlos olivine standard (11). In the following diagrams we use only high-precision data.

We have analyzed nearly 17,000 grains of olivine phenocrysts from 229 samples of tholeiitic to transitional compositions covering MORBs (40 samples) from Mid-Atlantic Ridge, East Pacific Rise, South-East Indian Ridge, and Knipovich Ridge; OIBs (138 samples) from Hawaiian Islands and Emperor Seamounts, Canary Islands, Reunion, Azores, and Iceland; LIPs (36 samples) from Ontong Java Plateau, Siberia, Emeishan, Karoo, Afar, and North Atlantic Province; komatiites and associated picrites (15 samples) from the Archean Abitibi greenstone belt in Canada and the Belingwe belt in Zimbabwe and South Africa; Proterozoic komatiitic basalts from Gilmour Island, Canada; and komatiites and picrites from Gorgona Island, Colombia. Most samples are picrites or olivine basalts containing large amounts of fresh, high-magnesium olivine phenocrysts. The samples are subdivided into four groups: (i) MORB; (ii) within plate magmas (WPM, magmas erupted far from plate boundaries) forming OIB emplaced over thin lithosphere (<70 km thick), WPM-THIN; (iii) WPM (OIB and LIP) emplaced over thick lithosphere (>70 km thick), WPM-THICK; and (iv) komatiites and associated magmas, KOMATIITES. Details of sample locations, references for sample descriptions, and their group correspondence are presented in table S2a.

The most-magnesian olivine compositions (defined by olivines phenocrysts with Fo within 1 mol % from a maximum Fo) for each specimen were averaged (table S2a) for the plots shown in Fig. 1. Individual olivine analyses are presented on fig. S4 and tables S2, c to f.

In addition to Mn and Ni concentrations, which strongly correlate with Fo (Fig. 1, A and

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C), we also plot Mn/Fe and Ni versus Mg/Fe ratios (Fig. 1, B and D). These ratios do not vary significantly with olivine fractionation (see model curves Frac 1 and Frac 2) but nevertheless range considerably (Fig. 1, B and D). Most olivine phenocrysts from MORBs and many from komatiites have Mn and Ni contents similar to those of peridotite-derived melts. In contrast, most olivines from the WPM-THICK group are significantly depleted in Mn and enriched in Ni. Their concentrations are not compatible with the melting products of common peridotites. The olivines from the WPM-THIN group have intermediate Mn and Ni contents.

Concentrations of Ca also provide some discrimination in spite of the greater overlap. Most olivines from the WPM-THICK group are too low in Ca to have precipitated from peridotite-derived melts (shown as experimental-based model compositions and fractionation trajectories, Fig. 1E).

Chromium is strongly controlled by garnet and spinel in peridotites and thus might be useful to decipher products of high-degree melting of

peridotite, which leave residuals (restites) free of Cr-rich phases (12). Olivines from Archean komatiites have the highest Cr values and match compositions of olivines from a spinel- and garnet-free refractory restite (Fig. 1F). They could, therefore, be derived directly from high-degree melting of peridotite. In the other groups of olivines, Cr is markedly lower than expected in equilibrium with peridotite at high pressures (see experimental data on lherzolite melting, Fig. 1F). The lowest Cr contents are found in MORB olivines, indicative of residual Cr spinel.

Cobalt (Fig. 2A) shows nearly uniform correlation with Fo for all rock groups, with possibly only minor (around 5%) relative enrichment in WPM-THICK and WPM-THIN over MORB (estimated from group average Co/Fe of table S2a). Decoupling of Co and Ni yields Ni/Co ratios of many WPM-THICK olivines that are unusually high for the Earth mantle (Fig. 2B).

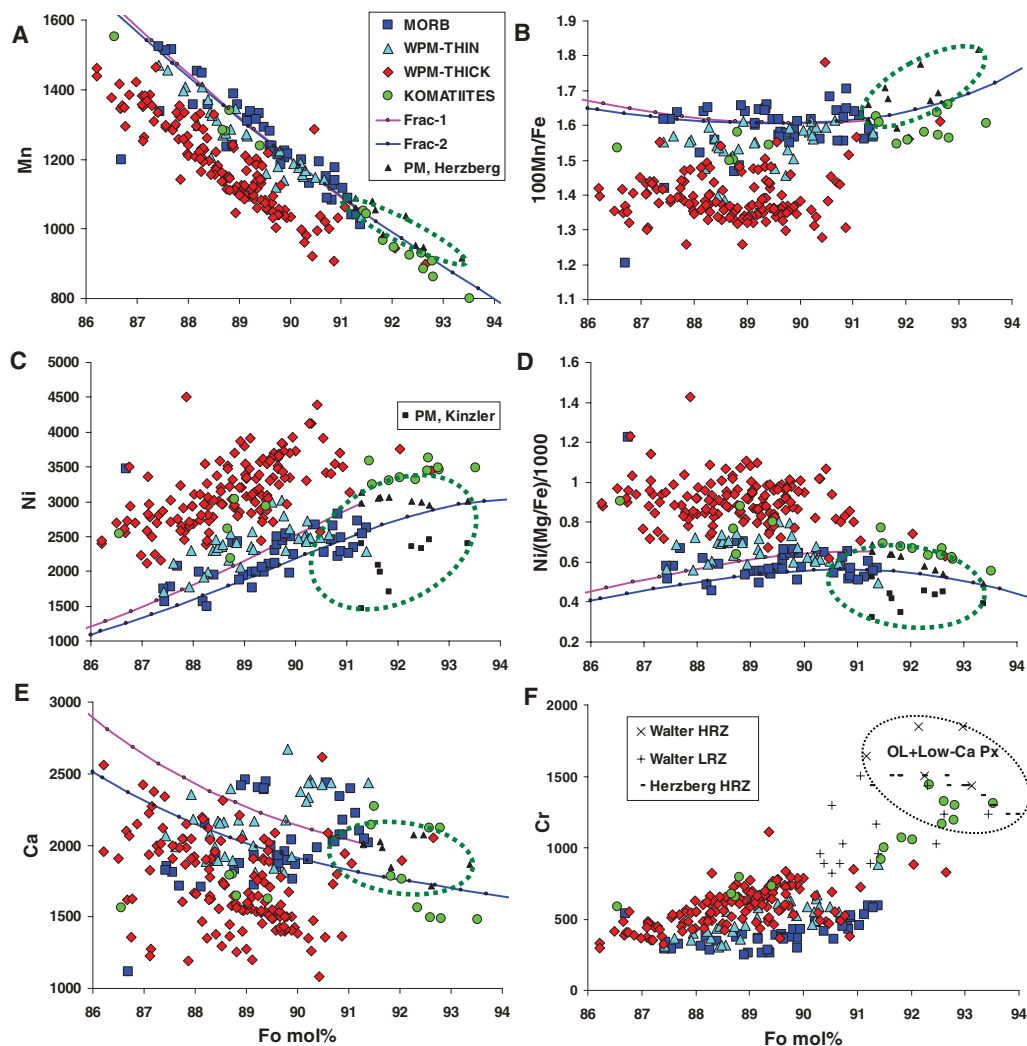
Mn/Fe is the parameter least dependent on olivine fractionation (as shown by the model fractionation curves in Fig. 3). Thus, it is diagnostic of parental magma compositional differ-

ences. There is a significant negative correlation of Ni/Mg versus Mn/Fe (linear correlation coefficient  $r$  is 0.66 for 238 samples) in spite of strong dependence of Ni/Mg on the degree of olivine fractionation (see fractionation trajectories in Fig. 3A). This correlation improves ( $r = 0.88$  for 103 samples) for the subset of olivines with a narrower range of Fo contents (Fo89 to Fo91). MORB olivines are the lowest in Ni and highest in Mn, whereas olivines from the WPM-THICK group are the highest in Ni and lowest in Mn, with olivines from the WPM-THIN group being intermediate.

To minimize the effects of olivine fractionation, we show parameters Ni/(Fe/Mg) and Ca/Fe in Fig. 3, B and C. This procedure also reduces the scatter in the ordinate significantly, thus highlighting the differences between geodynamic settings.

**Fate of recycled oceanic crust.** In subduction at  $P > 2.5$  GPa, the basaltic and gabbroic portions of the oceanic crust are transformed completely to eclogite (clinopyroxene and garnet) with a free SiO<sub>2</sub> phase (13–15). Unless silica has been

**Fig. 1.** (A to F) Average compositions of the most highly magnesian olivine phenocrysts in each sample. Concentrations and their ratios are given in ppm versus forsterite content of olivine in mol %. Olivine group names are as defined in text. PM, Herzberg indicates compositions of olivine in equilibrium at 0.1 MPa with melt originally generated at 3.0 to 5.0 GPa from fertile peridotite (12), calculated by Petrolog software (41) for oxygen fugacity corresponding to quartz-fayalite-magnetite (QFM) buffer using the Herzberg model (4), PM, Kinzler, olivine compositions similar to PM, Herzberg but with Ni calculated by using Ni partitioning between olivine and melt from Kinzler *et al.* (28). Frac 1 is the trend of olivine composition during fractional crystallization from a melt derived from fertile peridotite at 3 GPa and 1515°C (12). Fractionation of olivine modeled up to 20% for oxygen fugacity corresponding to QFM buffer using the Herzberg model (4). Frac 2 is similar to Frac 1 but calculated for 35% crystallization of melt derived at 4.0 GPa and 1630°C (12). Green ellipse indicates field of olivine compositions compatible with peridotitic source. In (F), HRZ, Herzberg stands for calculated compositions of olivine from spinel- and garnet-free harzburgite restite using (4); LHRZ, Walter and HRZ, Walter indicate experimental olivines from lherzolite- and garnet-free harzburgite residual assemblages, respectively, produced by high-pressure melting of fertile peridotite (12). Black ellipse marked "ol + low Ca-Px" indicates field of olivine compositions from refractory garnet- and spinel-free assemblage of olivine and low-Ca pyroxene.



removed during subduction (16), this combination will also be the relevant assemblage during recycling to the upper mantle (17).

In the ascending mantle (e.g., a mantle plume or upstream flow of convecting mantle), the silica-oversaturated eclogite starts melting at higher pressures than the peridotite and produces high silica melt (18, 19). This melt reacts with olivine from peridotite, producing pyroxenes and garnet (5, 8, 19). Previous studies have envisioned that this reaction creates a refertilized peridotite enriched in pyroxene (19, 20). This conclusion would predict variable mixing proportions of individual ingredients (eclogite-derived high-Si melt and peridotite) that are

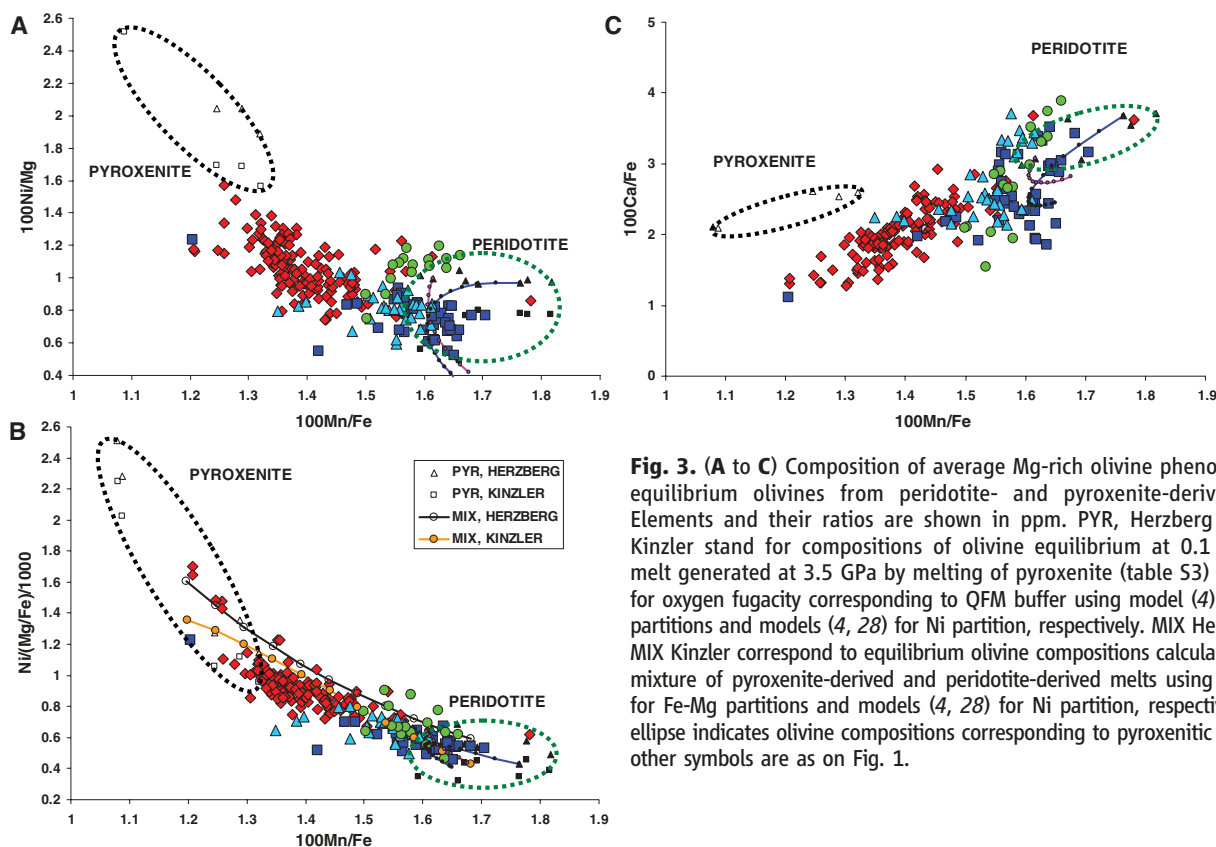
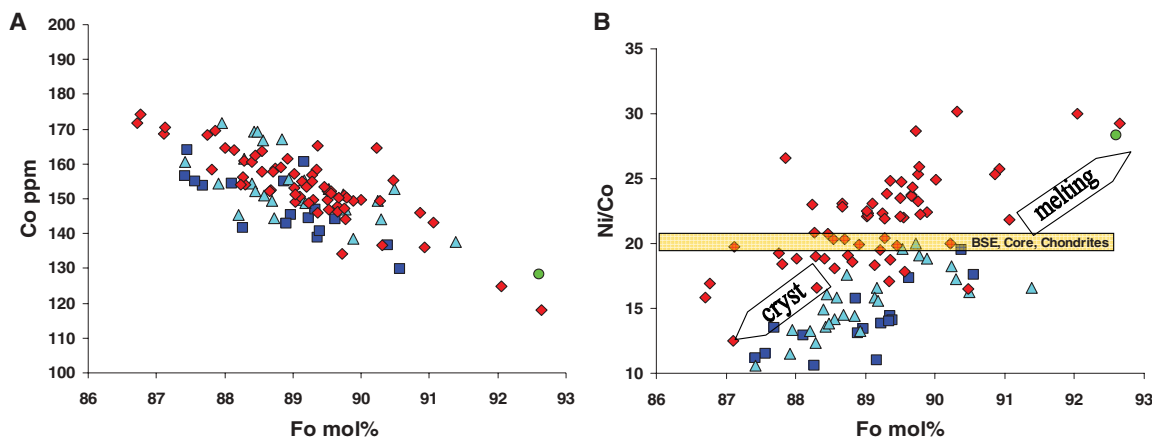
drastically different in composition. Melting such variable source compositions would create highly nonlinear correlations of  $^{187}\text{Os}/^{188}\text{Os}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  isotope ratios in the melts, and this is contradicted by the strongly linear correlations observed in Hawaiian basalts (21, 22), which are thought to have a significant eclogite component (3, 5).

However, it has been shown experimentally (23) and proposed on the basis of Korzinskii's theory (24) that, under conditions of local equilibrium, the reaction between high-Si eclogite-derived melt and peridotite produces an olivine-free lithology enriched in pyroxene (5). This fundamentally differs from a partial reaction (19, 20) because it leads to a stable pyroxenite lithology (hybrid py-

roxenite) generated by roughly fixed proportions of high-Si melt and peridotite [constrained by reaction stoichiometry between 40 and 60 weight % (wt %) of melt (5)] irrespective of the initial proportions of the reaction ingredients. Consequently, the hybrid pyroxenite has nearly uniform chemical and isotopic composition, thus constituting a single mixing endmember. Binary mixing of melts derived from peridotite and this pyroxenite leads to near-linear  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $^{87}\text{Os}/^{188}\text{Os}$  trends (5).

Other predicted geochemical consequences of replacement of olivine by pyroxene are a significant decrease of the bulk distribution coefficient between crystals and melt ( $K_d$ ) for Ni

**Fig. 2. (A and B)** Cobalt and nickel to cobalt ratio versus Fo of average Mg-rich olivine phenocrysts. Pink band at Ni/Co  $20 \pm 1$  represents estimated values for bulk silicate Earth (BSE), core, and chondrites (39). Arrows indicate trend of olivine compositions due to the mantle melting (melting) and magma crystallization (cryst). All other symbols are as on Fig. 1.



**Fig. 3. (A to C)** Composition of average Mg-rich olivine phenocrysts and equilibrium olivines from peridotite- and pyroxenite-derived melts. Elements and their ratios are shown in ppm. PYR, Herzberg and PYR, Kinzler stand for compositions of olivine equilibrium at 0.1 MPa with melt generated at 3.5 GPa by melting of pyroxenite (table S3) calculated for oxygen fugacity corresponding to QFM buffer using model (4) for Fe-Mg partitions and models (4, 28) for Ni partition, respectively. MIX Herzberg and MIX Kinzler correspond to equilibrium olivine compositions calculated for the mixture of pyroxenite-derived and peridotite-derived melts using model (4) for Fe-Mg partitions and models (4, 28) for Ni partition, respectively. Black ellipse indicates olivine compositions corresponding to pyroxenitic source. All other symbols are as on Fig. 1.

(5, 6) and a decrease in the ratio of the bulk coefficients of Mn and Fe (7). These features occur because olivine is the major silicate phase in peridotite concentrating Ni and the only silicate phase in peridotite having  $K_d$  for Fe greater than  $K_d$  for Mn (7). The bulk  $K_d$  changes will increase Ni and lower the Mn/Fe ratio of pyroxenite-derived melt compared with peridotite-derived melts. In addition, melting of pyroxenite yields lower Ca compared with peridotite (8). Additional predicted differences are higher melt fractions for hybrid pyroxenite than peridotite and higher Si and lower Mg in pyroxenite-derived melt (5, 25, 26).

These predictions were tested by experimental melting of a model hybrid pyroxenite (5). Experiments were run at  $P = 3.5$  GPa and temperatures between  $1400^\circ$  and  $1570^\circ\text{C}$  in a conventional 1.27-cm piston-cylinder apparatus at the Australian National University (10, 19). These results, together with published experimental data for melting of peridotite (12), confirm the Ni and Mn relationships as well as melting rates predicted above (table S3 and fig. S5). From these data, we calculate equilibrium olivine compositions at low pressure (4, 27, 28) in order to compare them with the natural phenocryst data (Fig. 3). We used these results to estimate mixing proportions of melts derived from the two end-member sources for the olivine data sets representing different geodynamic settings. The end-member melt compositions were calculated from averaging experimental data on melting of pyroxenite and peridotite (10).

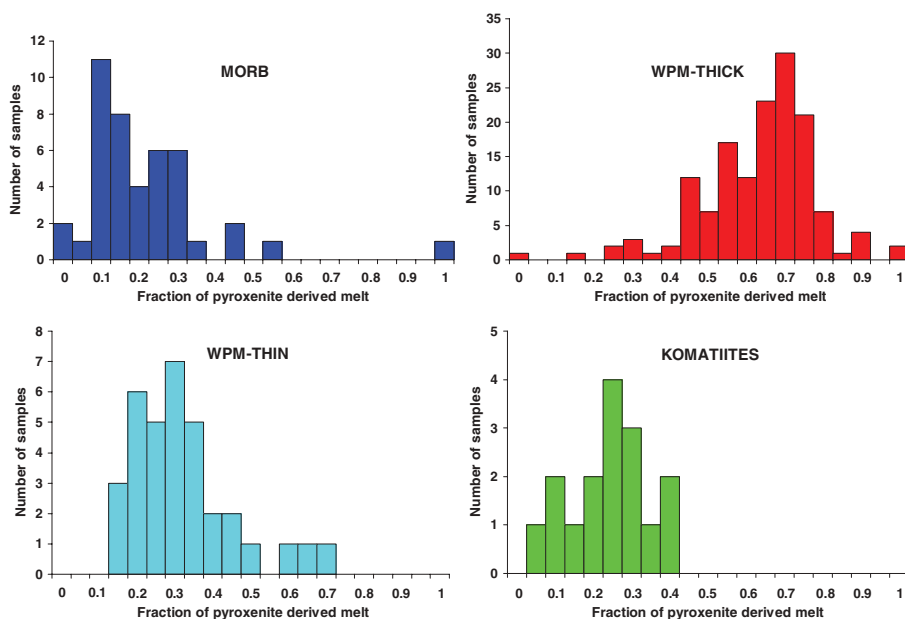
**Quantitative estimates.** We assumed mixtures (in 10% intervals) of the end-member melts (10) and calculated the composition of equilibrium olivines. The calculated mixing trajectories for the two different models for Ni partitioning

are consistent with natural olivine data (Fig. 3B). The relation between Mn/Fe of modeled olivines and mixing proportions (10) was used to compute the amount of pyroxenite-derived component for individual samples (Fig. 4). Olivines from the WPM-THICK group of basalts yield an average of  $61 \pm 16\%$  (standard deviation) pyroxenite-derived component, similar to results derived from Ni contents in Hawaiian melt inclusions and olivines only (5). The olivines from some continental LIPs (specific suits from Siberia and Karoo) indicate almost pure pyroxenitic sources. Corresponding results for the other groups are for WPM-THIN,  $30 \pm 13\%$ ; for Archean komatiites,  $21 \pm 10\%$ ; and for MORB, excluding one unusual sample from the Southern Atlantic (see below),  $17 \pm 12\%$  [similar to predictions of (25)]. Because of the uncertainties involved in estimating the end-member compositions, the differences between groups are better constrained than the absolute numbers. Although MORBs contain the lowest proportion of pyroxenite-derived melt, the spread of MORB data is significant, and many samples do contain substantial amounts of pyroxenite-derived component [the extreme is the enriched in silica MORB sample from the Southern Atlantic (29) with 100% pyroxenite-derived component]. The calculations show that the Archean komatiites contain a significant amount of pyroxenitic component (maximum of 30% for samples from Canada and Belingwe), although the largest amount is in Proterozoic komatiitic basalts from Gilmour Island, Canada (up to almost 40%). From these calculations, an estimate of the amount of recycled oceanic crust (10) yields 4% for MORB, 11% for WPM-THIN group, 16% for WPM-THICK group, and around 13% for Archean komatiites. The highest estimate of

the amount of recycled oceanic crust (10) yield Ontong Java high-Mg lavas: 13% to 28%.

**Silica-undersaturated basalts.** Most of the magmas analyzed in this study are silica saturated (tholeiites or transitional). Only a few samples are moderately silica-undersaturated alkali magmas (e.g., Azores and Afar). Our database is representative of the normal oceanic crust, several of the world's major suites of flood basalts (LIPs), several of the major modern mantle plumes (30), and some komatiites. The strongly silica-undersaturated associations not covered here include continental rift basalts, many smaller ocean islands consisting mostly of alkali lavas, and also some larger-flux plumes (30) such as Pitcairn, Tahiti, and Cape Verde islands. Why are such basalts that are highly enriched in incompatible elements, and therefore presumably generated by very low degrees of melting, nearly always undersaturated in silica? This observation appears to contradict our model; one would expect that silica-saturated melts generated from hybrid pyroxenites should be prevalent, especially at very low melt fractions. There are several possible explanations. (i) A volatile (mostly  $\text{CO}_2$ -) triggered melting of peridotite may be the dominant mechanism forming strongly silica-undersaturated alkaline magmas at temperatures lower than hybrid pyroxenite melts (31). (ii) Low-degree melts of silica-saturated eclogite may be retained in the source because of their high viscosity, thus preventing production of the hybrid pyroxenite (5). (iii) Melting of hybrid pyroxenite at the contact with peridotite may produce low-degree, silica-undersaturated melts at lower temperatures than melting of hybrid pyroxenite itself (8). (iv) Melting of bimineraleclogites (no free silica phase) formed from silica-undersaturated recycled crust produces undersaturated alkaline magmas (16).

**What controls the amount of pyroxenite-derived melt?** By following the method outlined above, we estimated the proportions of melt derived from pyroxenite and peridotite for each parental magma. These proportions depend on several interrelated parameters, namely the thickness of lithosphere, the potential mantle temperature ( $T_p$ ) (32), and the amount of recycled crust in the upwelling mantle (Fig. 5). Because at the same  $T_p$  pyroxenite melts at higher pressure than peridotite (26), a thick lithosphere (which imposes a high lower limit on the depth of melting) will suppress low-depth peridotite melting and therefore favor a high proportion of pyroxenite-derived melts (33, 34). The extreme case is found in some continental flood basalts (specific suites of Siberia and Karoo at table S2a) where the amount of pyroxenite-derived melt is nearly 100%. In such a case, the amount of recycled material cannot be estimated because the peridotitic component contributes no melt. In contrast, a thin lithosphere (MORB, Iceland, Azores, and Detroit seamount) favors a higher proportion of peridotite-derived melt because of the increasing degree of melting of peridotite at shallower depths.



**Fig. 4.** Estimated amounts of pyroxenite-derived component in the parental melt for 229 samples of four different groups.

A high  $T_p$  is an important condition to maintain sufficient buoyancy of mantle plumes or any other upstream mantle flow, and this buoyancy limits the amount of dense eclogite that they can carry (35). Also, a high  $T_p$  affects mostly the proportion of peridotite-derived melt because fractional melting imposes a rather stringent upper limit to further melting at high melt fractions (36). High melt fractions are restricted to the eclogite and hybrid pyroxenite assemblages (26) (fig. S5). The peridotite assemblage produces lower melt fractions than pyroxenite (fig. S5) or eclogite (19, 25) at any given temperature and pressure, and its actual extent of melting therefore depends strongly on the specific  $T_p$ .

Lastly, why are the proportions of recycled component lower beneath mid-ocean ridges than in thick-lithosphere settings? We suggest several explanations: (i) Relatively low amount of dense recycled component in MORBs is limited by their  $T_p$ , which is too low to carry more. (ii) For statistical reasons, plumes are more likely to encounter more-common thick lithosphere than less-common thin lithosphere and few plumes impinge directly on ridges, so we are forced to deal with very-small-number statistics. (iii) Detroit seamount represents one case where a (Hawaiian) plume has encountered thin lithosphere and where our results do indicate a high fraction of recycled crust, similar to those found on the island of Hawaii on the thick lithosphere.

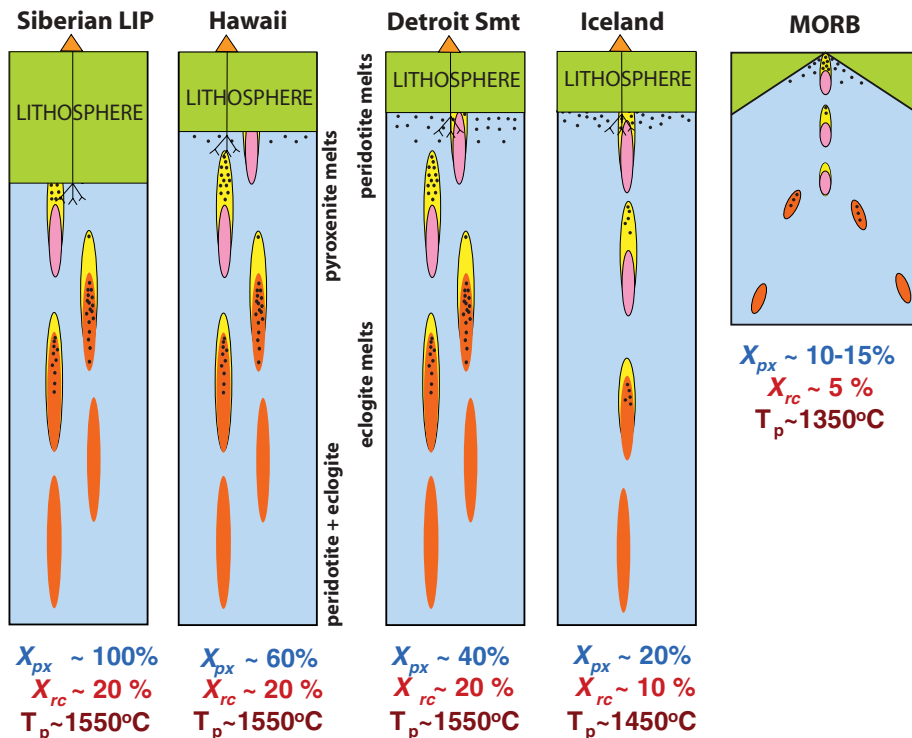
This amount is significantly higher than for Iceland, probably reflecting the effect of a higher  $T_p$  of the Hawaiian plume (4). (iv) The surface expression of a plume emplaced under thick lithosphere requires high  $T_p$ , which is necessary for carrying a significant amount of recycled crust (35), allowing melts to form at higher pressures than for ordinary peridotite (5, 25, 26), and melting a peridotite at higher pressures [e.g., komatiites (37)].

#### Heterogeneous versus homogeneous mantle.

The model presented here assumes that the recycled crustal component was not fully mixed with peridotite during subduction and mantle convection and thus that the formation of the olivine-free hybrid lithology may take place. On the other hand, homogenization of crustal material within the peridotite mantle should create a range of ultramafic lithologies with variable amounts of olivine, similar to a model by Kelemen *et al.* (20). Under these circumstances, the major-element contents of partial melts will correspond to the eutectic-like composition, buffered by the peridotite assemblage, whereas the compatible trace elements (Ni and Mn) will be controlled by the bulk partition coefficients of this assemblage and thus by the amount of olivine and pyroxene present in the system. Therefore, the amount of recycled crust can still be estimated on the basis of these trace elements, but their abundances will no longer correlate with the

buffered major elements (Si, Ca, and Al). For Hawaiian basalts, such correlations with Ni are present (5), which requires a strongly heterogeneous source.

**Input from Earth's core?** The Ni excess in mantle olivines from Siberian LIP (38) and the elevated Fe/Mn ratios in Hawaiian lavas (7) have been explained by input from Earth's core to the sources of mantle plumes. This suggestion is consistent with  $^{186}\text{Os}/^{188}\text{Os}$  ratios for some Hawaiian and Gorgona lavas (39) but is contradicted by the fact that concentrations of highly siderophile (Pt) and chalcophile (Cu) elements reported for Hawaiian basalts are not affected by this process (5). Our olivine data provide strong arguments against any notable core contribution to Ni or Fe excess in the sources of mantle-derived magmas. Cobalt does not show significant excess in olivines (Fig. 2 and fig. S3) and is effectively decoupled from Ni. As a result, the Ni/Co ratio in most Ni-rich mantle plume olivines exceeds 30 at the typical mantle Fo range of 89 to 91 (Fig. 2). This is not expected from a core contribution, because Ni/Co ratios for both mantle and core are almost equal and close to the chondritic value of about 20 (40). In addition, Ca is significantly depleted in many high-Ni and low-Mn olivines from the WPM-THICK group, which cannot be explained by core contribution. Lastly, the olivines from Gorgona komatiites, which do show significant excess in  $^{186}\text{Os}$  (39), do not indicate large anomalies in Ni and Mn/Fe, whereas Koolau lavas with the highest Ni excess and lowest Mn/Fe ratio in olivines do not show significant elevations in  $^{186}\text{Os}/^{188}\text{Os}$  ratios (39). This suggests complete decoupling of these potentially strong indicators of core-mantle exchange.



**Fig. 5.** Schematic model illustrating interplay between amount of recycled crust, thickness of lithosphere, and  $T_p$ . Blue, upwelling peridotitic mantle; red, recycled oceanic crust (eclogite with free  $\text{SiO}_2$  phase); black dots, melting; yellow, reaction zones forming hybrid pyroxenite; pink, refractory restite after eclogite melting; and green, lithosphere.  $X_{px}$ , amount of pyroxenite derived melt in the mixture with peridotite-derived melt, and  $X_{rc}$ , amount of recycled oceanic crust in the peridotitic mantle (42).

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43. We thank B. Schulz-Dobrick for supervising purchase and establishing electron microprobe facility in MPI; Hawaiian Scientific Drilling Project team, Koolau Scientific Drilling Project team, Ocean Drilling Program team, A. T. Anderson, E. A. Mathez, and N. Gitahi for providing samples; E. J. Jarosevich for supplying microprobe standards; A. Yasevich for sample preparations and N. Groschopf for maintaining the electron microprobe. The paper benefited from discussions with M. Hirschmann, P. Kelemen, C. Herzberg, B. McDonough, I. Ryabchikov, L. Kogarko, A. Kadik, E. Galimov, V. Batanova. Constructive reviews of C. Herzberg, and P. Kelemen improved the

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#### Supporting Online Material

www.sciencemag.org/cgi/content/full/1138113/DC1  
SOM Text  
Figs. S1 to S5  
Tables S1 to S4  
Data set

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# Genes Required for Mitotic Spindle Assembly in *Drosophila* S2 Cells

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The formation of a metaphase spindle, a bipolar microtubule array with centrally aligned chromosomes, is a prerequisite for the faithful segregation of a cell's genetic material. Using a full-genome RNA interference screen of *Drosophila* S2 cells, we identified about 200 genes that contribute to spindle assembly, more than half of which were unexpected. The screen, in combination with a variety of secondary assays, led to new insights into how spindle microtubules are generated; how centrosomes are positioned; and how centrioles, centrosomes, and kinetochores are assembled.

The diamond-shaped mitotic spindle has become one of the most widely recognized images in biology, emblematic of life's propagation through cell division. In higher eukaryotes, the process of spindle formation begins after nuclear envelope breakdown (NEB) when microtubules (MTs), generated both from centrosomes and from the vicinity of chromatin, are organized into a bipolar array (1–5). Sister chromatids bind to MTs emanating from opposite poles, are aligned in the middle of the bipolar MT network, and then ultimately separate and move apart during anaphase. Failures early in mitosis result in the formation of an abnormal metaphase

spindle, which can lead to mitotic delay and, potentially, chromosome missegregation during the ensuing anaphase.

To understand the mechanism of metaphase spindle assembly, it is critical to identify the proteins required for this process and then dissect how they function. Many mitotic proteins have been identified through genetic and RNAi screens (6–10), but the inventory is likely incomplete. Here, we present a genome-wide screen for mitotic spindle morphology in *Drosophila* S2 cells and the functional analysis of unexpected genes discovered through the screen.

**Identification of genes involved in metaphase spindle formation by high-throughput microscopy.** Because the percentage of S2 cells in mitosis is low (~1%), we conducted our RNAi screen in the presence of dsRNA (double-stranded RNA) to Cdc27 (a subunit of the anaphase-promoting complex) to delay anaphase and thereby increase the percentage of metaphase cells (~10% of the population). Thus, our screen was designed to investigate the assembly of the metaphase spindle, but not anaphase or cytokinesis. We also rescreened the

final hits without Cdc27 RNAi-induced mitotic arrest. The majority (88%) showed identical phenotypes, although a few genes only manifest clear phenotypes under conditions of mitotic arrest (table S1).

Using our custom, full-genome (14,425 genes) *Drosophila* RNAi library (11), we treated S2 cells with dsRNA for 4 days, conditions that generally reduce protein levels by >80% (12, 13). After dsRNA treatment, cells were fixed and stained for DNA,  $\gamma$ -tubulin, MT, and phosphohistone H3 (pH3) in 96 well plates, and about 40 sites per well were imaged by automated microscopy with a high numerical aperture air objective to obtain relatively high-resolution images of mitotic spindles (Fig. 1A). To reduce the complexity of this large amount of image data, a custom computer code was used to identify, crop, and arrange mitotic spindles into galleries, which were then blindly scored by an observer (Fig. 1B and fig. S1). In addition, computer algorithms measured eight parameters of spindle shape, as well as the intensity of  $\gamma$ -tubulin, cell number, and mitotic index (Fig. 1C) (11). More than 4,000,000 spindles were analyzed in this screen.

Before beginning this screen, we annotated 49 genes that produce mitotic defects in S2 cells (table S2). Of these 49 genes, 45 were identified as hits in the primary screen, indicating a high success rate of identifying mitotic phenotypes. However, our final list of genes should not be considered as a complete or universal inventory, because genes can be missed (particularly those with subtle phenotypes), and some phenotypes (or lack thereof) may be specific to S2 cells. False positives by off-target effects of dsRNA can occur in RNAi screens (14, 15), so precautions were taken to minimize gene overlap in the dsRNA design, and all unexpected hits were confirmed with another dsRNA that had no overlap with the first dsRNA (11). To learn more about the functions of interesting genes,

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Corrected 19 April 2007: Minor typographic errors have been corrected.



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## Supporting Online Material for

### **The Amount of Recycled Crust in Sources of Mantle-Derived Melts**

Alexander V. Sobolev, Albrecht W. Hofmann, Dmitry V. Kuzmin, Gregory M. Yaxley, Nicholas T. Arndt, Sun-Lin Chung, Leonid V. Danyushevsky, Tim Elliott, Frederick A. Frey, Michael O. Garcia, Andrey A. Gurenko, Vadim S. Kamenetsky, Andrew C. Kerr, Nadezhda A. Krivolutskaya, Vladimir V. Matvienkov, Igor K. Nikogosian, Alexander Rocholl, Ingvar A. Sigurdsson, Nadezhda M. Sushchevskaya, Mengist Teklay

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#### **This PDF file includes:**

Materials and Methods  
Figs. S1 to S5  
Tables S1, S2a, S2b, and S3  
References

**Other Supporting Online Material for this manuscript includes the following:**  
available at [www.sciencemag.org/cgi/content/full/1138113/DC1](http://www.sciencemag.org/cgi/content/full/1138113/DC1)

Table S2 (complete) as zipped file

# The Amount of Recycled Crust in Sources of Mantle-Derived Melts

A. V. Sobolev, A. W. Hofmann, D. V. Kuzmin, G. M. Yaxley, N. T. Arndt, et al.

## Supporting Online Material

### Methods and Samples

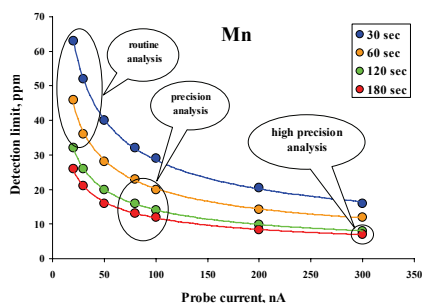
**Electron probe microanalysis.** Olivine phenocrysts were analyzed for Si, Fe, Mg, Ca, Ni, Mn, Cr, Al and Co with a Jeol Jxa 8200 electron probe at Max-Planck Institute of Chemistry. Each olivine grain was analyzed in the geometrical center at 20 kV accelerating voltage and probe current of 300 nA. A small olivine subset was analyzed at 30 kV and 200nA. Details of analytical conditions are presented in Table S1. These conditions have been found optimal to obtain best detection limits (see Fig S1). The intensity of the Co K $\alpha$  line was additionally corrected for overlap with the shoulder of FeK $\beta$  second order line using the following linear equation established by analyzing Fe bearing standards free of Co:

$$\text{CoO}_c = \text{CoO}_m - 0.0011 \times \text{FeO}_m - 0.013. \quad [\text{S1}]$$

Where  $\text{CoO}_c$  and  $\text{CoO}_m$ ,  $\text{FeO}_m$  are corrected and measured values correspondingly.

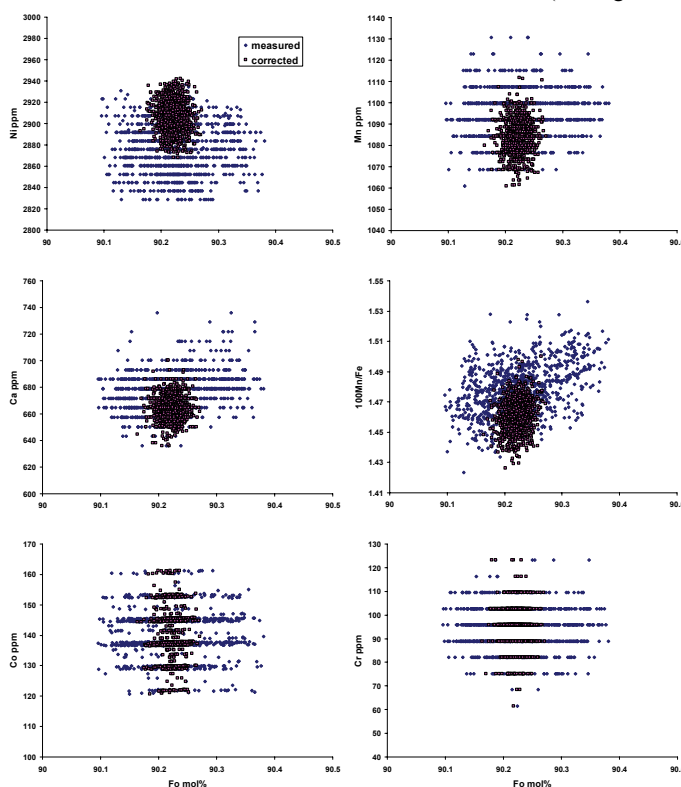
The constant in equation [S1] was obtained from analyses of San Carlos olivine standard with known Co content (Table S1). Note that the linear approximation of equation [S1] may not be valid for highly Mg (Fo>93) and low Mg (Fo<70) olivines.

In addition San Carlos olivine standard USNM 111312/444 (S1) was run (3 points per each 30-50 measurement points) as an



**Fig. S1.** Detection limit for Mn as function of probe current and peak counting time (in seconds, sec).

Detection limit was determined for San Carlos olivine standard with Jeol software using background statistics and 3 sigma criterion. Typical measurement conditions used in literature are marked as "routine analysis". Analysis at higher beam current and counting times published by (S3) is marked as "precision analysis". The analytical conditions used in this study are indicated as "high precision analysis".



**Fig. S2.** Reproducibility of San Carlos olivine standard analyzed three times every 30 to 50-th measurement points. Measured – represent uncorrected measurements, corrected – stands for measurements corrected for the calibration drift.

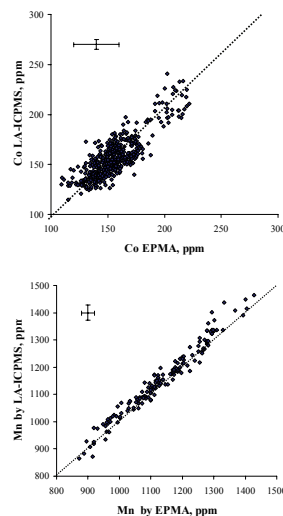
**Table S1.** Typical analytical conditions of electron probe microanalysis of olivine. Probe current 300 nA, acceleration voltage 20kV.

El	St	Cryst	Line	Peak	BG(+)	BG(-)	S.C.O.
Si	1	TAP	K $\alpha$	90	90	-	190740 *
Al	2	TAP	K $\alpha$	240	120	120	170 #
Fe	1	LIF	K $\alpha$	90	90	-	74230 *
Mn	3	LIFH	K $\alpha$	120	60	60	1084 *
Mg	1	TAP	K $\alpha$	90	45	45	298040 *
Ca	4	PETJ	K $\alpha$	120	60	60	665 #
Ni	5	LIF	K $\alpha$	150	70	70	2907 *
Co	6	LIFH	K $\alpha$	120	120	-	140 **
Cr	7	PETJ	K $\alpha$	120	60	60	95 #

**Notes for Table S1:** Standards (St) - 1- San-Carlos olivine USNM 111312/444 (S.C.O.); 2-pure Al $_2$ O $_3$ ; 3-Rhodonite; 4-Wollastonite; 5-pure NiO; 6- pure Co-metal; 7-pure Cr $_2$ O $_3$ . Peak and background (BG) counting time in seconds. \*-Accepted values for (S.C.O.) from ref (S1) in ppm, \*\*-value measured by LA-ICP MS.

unknown to monitor drifts in calibration and estimate accuracy of analysis. All measurements of Si, Fe, Mn, Ni, Ca and Al were corrected for deviation of this standard from the reference values (Table S1).

For trace elements, the above measurement conditions routinely yield detection limits of around 6-15 ppm based on 3 sigma criteria by Jeol standard software, and errors of 15-30 ppm for trace elements and 0.01 mol% for Fo content (2 standard errors) established by counting statistics and reproducibility of olivine standard (see Fig S1-S2). Precision and



**Fig. S3.** Comparison between Co and Mn concentrations in olivine measured by EPMA and LA-ICPMS. Error bar corresponds to  $\pm 2$  standard errors for both EPMA and LA-ICP MS

accuracy of Co and Mn measurements were independently checked by LA-ICP MS analysis using Thermo Finnigan Element 2 mass spectrometer and New Wave Up213 YAG laser (213 nm wavelength) setup. KL2-G (S4) and NIST 612 glasses were used as external standards and Ca in olivine as reference value. Figure S3 shows that both Co

Table S2f. These tables are included as separate spreadsheets together with Table S2a in the Excel file called Table S2. Table S2a is also placed in the end of this file. Tables S2c,d,e,f include concentrations of oxides and their relative standard deviations in % (signal counting statistics) for individual olivine grains. Each analytical point in addition to

**Melting of pyroxenites.** The model hybrid pyroxenite composition from Sobolev et al. (S3) (their Table S2, column 50%) was chosen for the high pressure experimental investigation. A synthetic starting material with this composition (Px-1) was prepared for the experiments by blending high purity oxides and carbonates under analytical grade

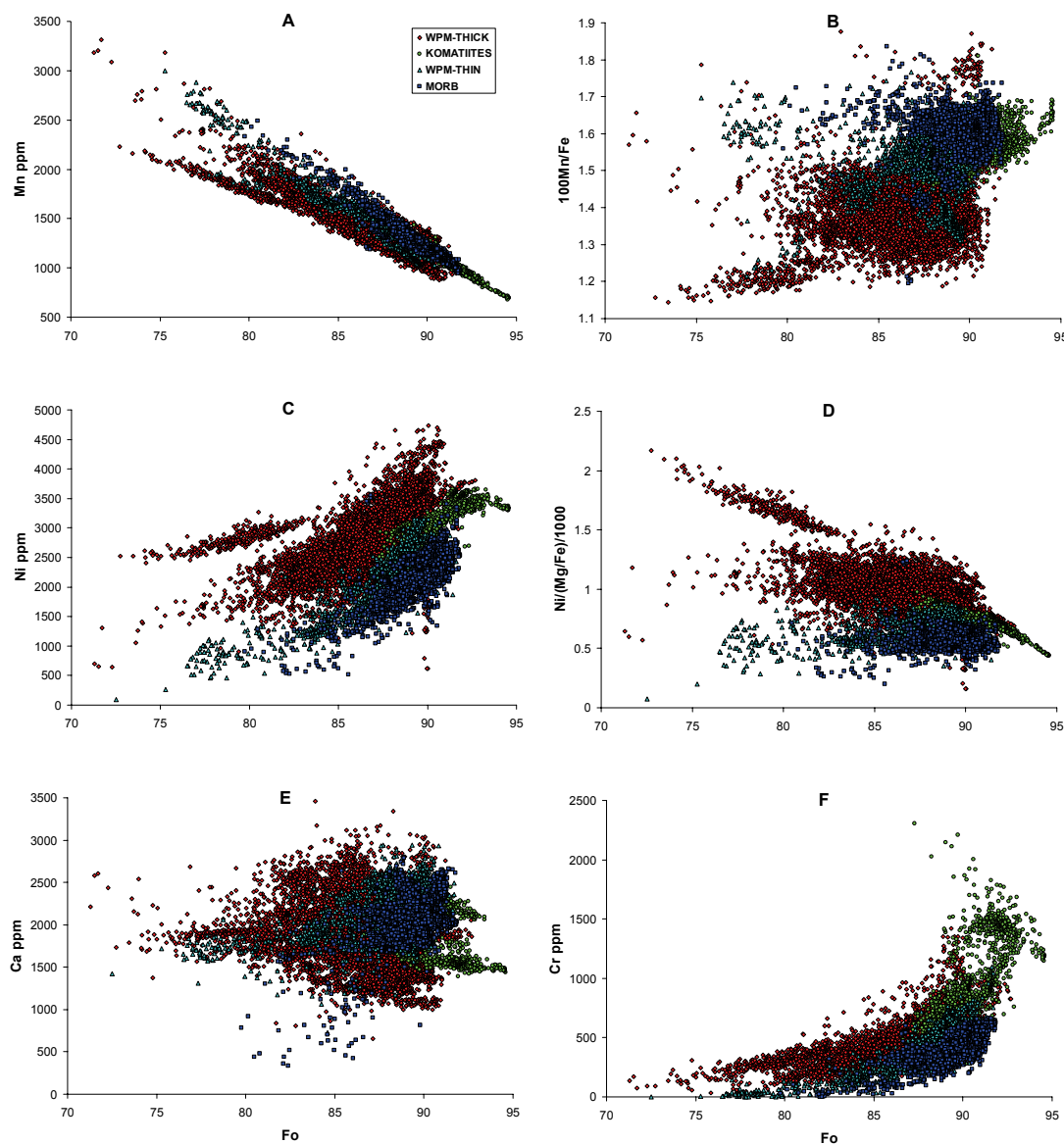


Fig. S4. Composition of olivine phenocrysts: all data from database. Group of MORB olivines shown on top of all groups.

and Mn concentrations measured by EPMA correspond to those analyzed by LA-ICP MS within 20 ppm (2 standard errors).

**Olivine database.** The averaged most highly magnesian olivines for each sample are presented in Table S2a. Individual olivine analyses are presented on Fig S4 and in Tables S2c,d,e,f separately for each group: MORB in Table S2c, WPM-THIN in Table S2d, WPM-THICK in Table S2e and KOMATIITES in

sample name has a unique number. Table S2a includes group title, sample name, reference for sample description (if published), information on locality, number of averaged high-Mg grains, forsterite content, element concentrations in ppm, standard errors of the mean in ppm, characteristic ratios of elements and calculated amount of pyroxenitic component ( $X_{px}$ , see below for explanation).

acetone for several hours, until the material was homogenous and very fine-grained. The resultant dried powder was then pelletised and fired for 12 hours at 1100°C to decarbonate and partially fuse the components. FeO was added after firing in the form of synthetic fayalite ( $Fe_2SiO_4$ ), and again blended under acetone. The final mixture was then dried overnight at 200°C and subsequently stored at 110°C. The actual composition of the Px-1

starting material was checked by electron probe microanalysis of quenched glass prepared by fusing the starting mix at 1300°C in an Ar atmosphere in a box furnace and quenching in water (Table S3).

High pressure experiments were run at a pressure of 3.5 GPa and temperatures between 1400-1570°C in a conventional 1.27cm piston-cylinder apparatus at the Australian National University. The Px-1 starting material ( $\approx 1$  mg) was loaded into graphite capsules, which were sealed by arc-welding in Pt outer capsules. The

capsules were placed centrally into NaCl-pyrex sleeves with MgO inserts, an assembly which requires no friction correction. Type B thermocouples ( $\text{Pt}_6\text{Rh}_{94}/\text{Pt}_{30}\text{Rh}_{70}$ ) were employed, with the thermocouple join placed within a fraction of a millimeter of the Pt capsule. Runs were brought to run temperature and pressure simultaneously. Temperature was controlled throughout by Eurotherm controllers attached to the thermocouple and is accurate to  $\pm 10^\circ\text{C}$ . Pressure is accurate to  $\pm 0.1$  GPa. Experiments were maintained at the

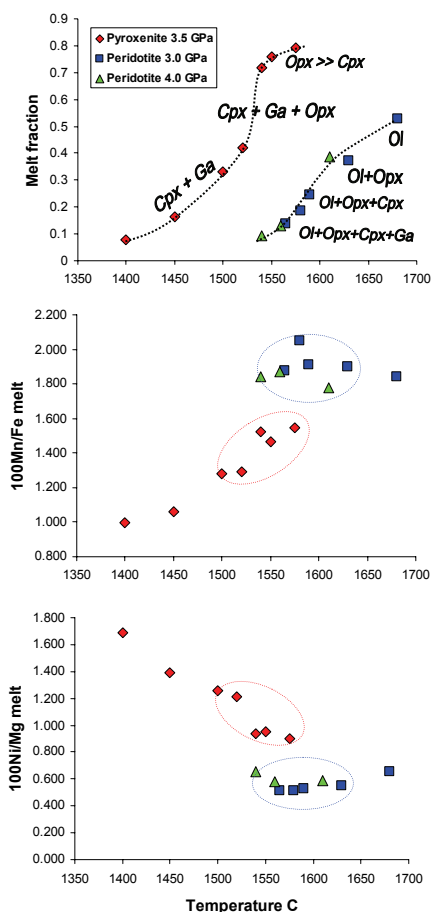
desired PT condition long enough to allow a close approach to equilibrium, and then quenched by terminating power to the furnace.

After recovery from the post-run assembly, the Pt capsules were mounted in 1 inch diameter epoxy buttons, sectioned longitudinally and polished, in preparation for analysis by scanning electron microscopy at the Australian National University and electron probe microanalysis at the Max Planck Institute for Chemistry in Mainz.

**Table S3.** Proportions of phases produced and melt compositions from melting of pyroxenite at 3.5 GPa and 1400-1575°C.

Run	DT	T	Phases proportions, wt%					Melt compositions in oxide wt%, Ni in ppm										
			N	h	°C	melt	Opx	Cpx	Ga	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O
C-2298	45	1400	8	0	79	14	60.70	2.76	14.40	6.80	0.068	4.44	7.71	2.34	0.70	0.07	453	
C-2317	48	1450	16	0	71	12	55.76	2.33	14.20	7.63	0.081	7.45	7.95	2.16	0.44	0.06	626	
C-2301	5	1500	33	0	60	6	53.30	1.58	14.21	8.25	0.106	10.67	8.61	1.75	0.13	0.09	810	
C-2330	5	1520	42	0	55	3	52.54	1.29	14.22	8.65	0.112	11.66	8.65	1.98	0.14	0.11	850	
C-2332	5	1540	72	24	4	0	52.00	0.86	13.55	8.19	0.125	14.08	9.00	1.80	0.08	0.20	796	
C-2318	6	1550	76	18	7	0	52.38	0.83	13.39	7.95	0.117	14.67	8.77	1.74	0.09	0.20	843	
C-2319	5	1575	79	20	0	0	52.20	0.79	13.07	8.11	0.126	15.44	8.52	1.67	0.07	0.22	841	
<b>Px-1, Bulk composition</b>								<b>52.67</b>	<b>0.64</b>	<b>11.26</b>	<b>7.55</b>	<b>0.119</b>	<b>18.48</b>	<b>7.05</b>	<b>1.52</b>	<b>0.06</b>	<b>0.25</b>	<b>1000</b>
<b>Peridotite-derived endmember</b>								<b>46.79</b>	<b>0.85</b>	<b>11.50</b>	<b>9.68</b>	<b>0.185</b>	<b>19.07</b>	<b>10.00</b>	<b>0.84</b>	<b>0.47</b>	<b>0.39</b>	<b>642</b>
<b>Pyroxenite-derived endmember</b>								<b>52.56</b>	<b>1.07</b>	<b>13.71</b>	<b>8.24</b>	<b>0.117</b>	<b>13.32</b>	<b>8.72</b>	<b>1.79</b>	<b>0.10</b>	<b>0.16</b>	<b>830</b>

**Note for Table S3:** Melt, Opx, Cpx, Ga, - proportions of melt, orthopyroxene, clinopyroxene and garnet in experimental runs in mass fractions, calculated using list square method from bulk composition and compositions of phases. DT- run duration in hours, T- temperature in °C. The amount of Ni has been calculated using the method described in Sobolev et al, 2005 (S3). This method uses known phase proportions and the distribution of Ni between phases, to calculate Ni in the phases on the basis of mass balance and known bulk Ni content. The assumed bulk Ni concentrations are: 1000 ppm for pyroxenite (Px-1) and 1900 ppm for peridotite (S3).



### Pyroxenite-derived component in melts.

In order to estimate endmember compositions we have averaged compositions of melts of pyroxenite and peridotite (see Fig.S5). For pyroxenite we averaged all melt compositions except numbers C-2298 and C-2327 (Table S3) produced at low temperature and low percentage of melting. For peridotites we averaged 7 compositions of melts from runs 30.12, 30.07, 30.14, 30.1 (3.GPa), 40.06, 40.07, 40.05 (4.GPa) of Walter, 1998 (S2). The endmember compositions are shown in Table S3. We assumed that the mixtures consist of endmembers with similar temperatures, and we therefore excluded (from the averaging procedure) experimental melts with very low or very high temperatures (see Fig.S5).

The calculated endmember melts were then mixed (in 10% intervals) producing mixed melt compositions. For these melts the composition of equilibrium olivines were

calculated using Herzberg's model (S5). The calculated Mn/Fe ratios of the olivines and the amounts of pyroxenitic endmember ( $X_{px}$ ) yield a straight line:

$$X_{px} = 3.48 - 2.071 \times (100\text{Mn}/\text{Fe}) \quad [\text{S2}]$$

We used this equation to calculate the amount of pyroxenitic endmember for each average olivine composition in Table S2a.

The calculated endmember compositions are derived from experiments at high pressures and temperatures. Therefore, they are directly relevant primarily to thick-lithosphere settings, but we have applied them for all settings. Equivalent estimates for lower pressures and temperatures (1.0 GPa and 1300-1350°C) have been calculated using pMELTS (S6). This yielded similar degrees of melting and residual assemblages (50 to 60% of melting and low-Ca pyroxene dominated residue) as those derived from high pressure and temperature runs (Table S3). This gives us some confidence that the same Mn/Fe ratios can be applied to lower pressures, but experimental confirmation will be needed. Additional confidence in the results is derived from the observation that the extreme compositions observed in both MORB and WPM-thick settings match the calculated endmember compositions reasonably well.

**Fig. S5.** Melt fraction and composition of experimental melts of pyroxenite and fertile peridotite. Temperatures for experimental runs of fertile peridotite (S2) are extrapolated to 3.5 GPa by using slope 100C/GPa. Data for pyroxenite from Table S3. Crystalline phase coexisting with melt are: Ol - olivine; Opx - orthopyroxene; Cpx - clinopyroxene; Ga - garnet. Blue and red ellipses indicate analyses averaged to calculate peridotite-derived and pyroxenite-derived endmembers respectively.

**Oceanic crust in magma sources.** A final step is to estimate the actual amount of recycled oceanic crust ( $X_{crc}$ ) from the determined proportion of pyroxenite-derived melt. This quantity is linked to the proportion of hybrid pyroxenite-derived melt ( $X_{px}$ ) by the degree of melting of eclogite ( $F_e$ ), the amount of eclogite-derived melt needed to produce hybrid pyroxenite from peridotite ( $X_e$ ), and the degrees of melting of peridotite ( $F_{pe}$ ) and pyroxenite ( $F_{px}$ ) (S3):

$$X_{crc} = \frac{X_e}{F_e \left( \frac{1 - X_{px}}{X_{px}} \frac{F_{px}}{F_{pe}} + \frac{1 - F_e}{F_e} X_e + 1 \right)} \quad [S3]$$

Following assumption of Sobolev et al, (2005) (S3), we propose that the extent of melting of eclogites reaches a maximum of 50%. This limit is unlikely to be exceeded significantly during fractional melting, because this process removes Na (S7) and thereby renders the residue highly refractory. Therefore it is also not strongly dependent on the potential temperature of the rising mantle material. The amount of pyroxenite produced by reaction of the primary (eclogite-derived) melt with peridotite is prescribed by the stoichiometry of the reaction, which also yields proportions of the reactants of close to 50:50 (S3). For these, proportions, the amount of reaction pyroxenite equals the amount of initial eclogite. The melt fraction produced by the pyroxenite can be estimated at low pressure (1-2 GPa by modeling using pMELTS software (S6)), yielding a maximum of about 60 % for batch melting at 1 GPa and 1320 to 1350°C. At higher pressures (3.5 GPa) we must rely on the experimental data (Table S3, Fig. S5), which also yield a maximum melt fraction of about 50-60% for batch melting for 1540-1550°C.

These values will be significantly lower for fractional melting (S8), for the same reason as discussed for eclogite melting (i.e. early Na removal). Adopting an amount of 50 % pyroxenite melting will therefore result in a minimum estimate in the amount of recycled crust present. Finally, we estimate the melt fraction of peridotite using published models (S5) for MORB (10%), Iceland, as a proxy for the WPM-THIN group (20%), Hawaii, as proxy for the WPM-THICK group (10%) and Archaean komatiites -40% (S9). This yields the following average estimates for the amounts of recycled oceanic crust in the mantle sources: 4% for MORB, 11% for WPM-THIN group, 16% for WPM-THICK group, and around 13% for Archaean komatiites. According to (S10) the melt fraction of peridotite for Ontong Java high-Mg magmas corresponds to 30%, which yield the amount of recycled oceanic crust of 13-28% in the mantle source of these magmas.

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**Table S2b.** Locations (GPS coordinates) and description for Icelandic samples

Sample	Locality	Description	latitude	longitude
01-7	Reykjanes, Haleyjabunga	Glassy olivine phryic picrite	N63° 48.954'	W22° 38.646'
03-102	Reykjanes, Lagafell	Glassy olivine phryic picrite	N63° 52.906'	W22° 32.611'
03-106	Reykjanes, Haleyjabunga	Glassy olivine phryic picrite	N63° 48.873'	W22° 38.724'
01-15	Reykjanes, Sulur	Glassy olivine -phyric basalt	N63° 54.094'	W22° 32.558'
01-12	Reykjanes, Stapafell	Glassy olivine-plagioclase phryic picrite	N63° 54.601'	W22° 31.370'
01-8	Hengill, Midfell	Glassy olivine phryic picrite with rare plagioclase and clinopyroxene xenocrysts	N64° 10.140'	W21° 04.257'
01-10	Hengill, Midfell	Glassy olivine phryic picrite with rare plagioclase and clinopyroxene xenocrysts	N64° 10.103'	W21° 04.118'
01-19	Hengill, Maelifell	Glassy olivine phryic basalt with rare plagioclase and clinopyroxene xenocrysts	N64° 06.252'	W21° 10.744'
03-131	Kistufell	Glassy olivine phryic basalt	N64° 47.806'	W17° 13.745'
03-164	Kistufell	Glassy olivine phryic basalt	N64° 46.918'	W17° 12.713'
03-140	Kistufell	Glassy olivine phryic basalt	N64° 47.861'	W17° 12.202'
03-161	Kistufell	Glassy olivine phryic basalt	N64° 47.808'	W17° 13.957'
01-55	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 56.285'	W17° 05.047'
01-57-4	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 55.791'	W17° 04.463'
01-56-1	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 56.277'	W17° 05.374'
01-41	Theistareykir, Theistareykjahraun	Glassy olivine phryic basalt	N65° 57.547'	W17° 04.120'
01-44	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 56.068'	W17° 05.246'
01-56-2	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 56.277'	W17° 05.374'
01-54	Theistareykir, Laufrandhraun	Glassy olivine phryic picrite	N65° 56.281'	W17° 04.624'
01-51	Theistareykir, Langavittshraun	Glassy olivine phryic basalt with rare clinopyroxene phenocrysts	N65° 56.058'	W16° 52.282'
03-224	Snaefellsness, Enni	Glassy olivine phryic basalt	N64° 54.146'	W23° 45.776'
03-226	Snaefellsness, Sydri-Raudamelur	Olivine phryic basalt	N64° 52.296'	W22° 17.368'
03-220	Snaefellsness, Ytri-Raudamelur	Olivine phryic basalt	N64° 52.717'	W22° 20.860'

Table S2a. Average compositions of most Mg-rich olivine phenocrysts (see text of paper for definition).

GROUP	Geodynamic setting	References	Sample	Geographic	Locality	N	Fo	Al ppm	Fe ppm	Min ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm	Mg ppm	Fe ppm	Al ppm	Fe ppm	Mn ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm				
MORB		(S1)	113-16	Indian Ocean	105.22-0E-SEIR	24	88.49	270	79979	1233	296327	1935	2109			389	1,617	0.712	0.569	2,494	0.135	0.07	7	502	10	268	13	38	5
MORB		(S1)	145-3	Indian Ocean	116.72-0E-SEIR	9	89.90	298	77032	1224	286238	1761	2500			369	1,590	0.838	0.646	2,265	0.191	0.05	4	422	8	234	9	27	8
MORB		unpb	30-29	Indian Ocean	124.28-0E-SEIR	10	89.60	236	79185	1280	296874	1843	2115			390	1,617	0.712	0.564	2,238	0.135	0.10	10	724	14	499	20	71	9
MORB		(S1)	144-13	Indian Ocean	115.21-0E-SEIR	29	88.16	269	82315	1316	394471	1921	1983			363	1,599	0.677	0.557	2,334	0.171	0.05	6	377	8	265	20	39	5
MORB		unpb	18-10	Indian Ocean	126.74-0E-SEIR	24	89.60	272	78878	1239	295854	1921	2303			372	1,571	0.778	0.610	2,435	0.06	0.06	4	438	9	280	12	38	7
MORB		(S1)	68-1	Indian Ocean	88.92-0E-SEIR	11	88.89	319	76966	1205	293733	2036	1972			299	1,566	0.682	0.510	2,446	0.239	0.05	10	417	8	287	10	29	4
MORB		(S1)	88-22	Indian Ocean	96.93-0E-SEIR	10	86.18	213	86827	1453	287837	2153	1572			380	1,639	0.546	0.484	2,429	0.088	0.11	8	785	14	526	10	19	7
MORB		unpb	5-2	Indian Ocean	126.52-0E-SEIR	33	88.40	257	72888	1238	294600	1875	2383			364	1,582	0.809	0.633	2,395	0.207	0.04	3	307	6	235	3	17	2
MORB		unpb	16-19	Indian Ocean	127.58-0E-SEIR	18	88.40	235	87179	1357	289113	1920	1953			346	1,557	0.676	0.589	2,202	0.258	0.15	5	1112	18	675	21	48	6
MORB		unpb	JPE-1	Pacific Ocean	Garrett F.Z.	32	90.82	395	69627	1084	299751	2071	2817			480	1,557	0.940	0.654	2,975	0.259	0.02	10	171	3	169	3	38	11
MORB		(S2)	SO-1	Pacific Ocean	Siqueleros F.Z.	89	90.38	377	72454	1131	286261	1985	2661	137		413	1,561	0.898	0.651	2,739	0.250	0.03	4	191	3	201	2	15	1
MORB		(S1)	FZ 649-5(i)	Atlantic Ocean	9 N MAR	36	89.75	301	76919	1213	293900	1953	2444			374	1,577	0.834	0.642	2,539	0.216	0.04	5	309	4	200	5	15	4
MORB		(S3)	FZ 649-5(ii)	Atlantic Ocean	9 N MAR	28	88.76	302	84161	1235	289205	1832	2410			371	1,467	0.833	0.701	2,177	0.444	0.04	7	276	6	173	14	30	6
MORB		(S3)	649(i)	Atlantic Ocean	9 N MAR	12	89.63	314	78044	1222	293379	1929	2501	144		360	1,566	0.853	0.665	2,472	0.239	0.05	7	410	6	334	18	20	2
MORB		(S3)	649(ii)	Atlantic Ocean	9 N MAR	10	88.87	310	83452	1240	298834	1875	2434	155	390	1,486	0.840	0.840	0.701	2,247	0.406	0.08	9	639	10	585	20	55	8
MORB		unpb	All-60_Sun9_D9_9-20	N. Atlantic	43N MAR	88	88.81	332	76416	1184	293134	1896	2543			430	1,549	0.868	0.663	2,481	0.274	0.02	2	152	2	103	2	11	3
MORB		unpb	All-32-11-92	N. Atlantic	43N MAR	92	90.70	242	70116	1151	297618	2032	2461			502	1,642	0.827	0.580	2,889	0.268	0.01	5	109	2	97	9	7	5
MORB		unpb	All-32-12-8(i)	N. Atlantic	43N MAR	74	90.83	240	69316	1124	298835	2340	2221			428	1,622	0.743	0.515	3,376	0.124	0.03	3	200	3	134	14	19	10
MORB		unpb	All-32-12-8(ii)	N. Atlantic	43N MAR	13	90.88	252	68934	1175	298950	2174	2281			510	1,705	0.766	0.528	3,153	-0.048	0.01	3	57	2	166	33	10	3
MORB		unpb	All-32-12-7	N. Atlantic	43N MAR	85	91.13	254	66863	1089	298972	2221	2339			531	1,629	0.782	0.523	3,321	0.110	0.03	4	201	4	132	27	18	10
MORB		(S4)	All-32-12-7	N. Atlantic	43N MAR	55	91.03	213	67956	1115	300251	2393	2252			453	1,641	0.750	0.510	3,322	0.084	0.03	4	211	4	152	22	11	10
MORB		(S5)	All-32-11-78	N. Atlantic	43N MAR	68	90.88	260	69165	1140	299897	2072	2479			521	1,648	0.827	0.572	2,996	0.270	0.03	5	202	3	126	12	15	6
MORB		(S5)	CH31/D08.685	N. Atlantic	Famous	17	91.21	306	66586	1039	300597	2100	2721			592	1,561	0.905	0.603	3,154	0.251	0.09	9	670	12	371	8	81	8
MORB		(S5)	CH31/D08.136-02	N. Atlantic	Famous	19	91.38	298	64581	1012	297930	2019	2620			595	1,567	0.879	0.568	3,127	0.237	0.07	10	523	9	285	7	89	8
MORB		(S5)	ARP73.10.03	N. Atlantic	Famous	52	91.30	295	65425	1039	298985	2046	2535			582	1,589	0.848	0.555	3,126	0.193	0.05	9	345	5	180	5	31	10
MORB		(S6)	AG32-4-68	S. Atlantic	Bouvet	44	90.72	319	69996	1088	297891	2024	2653			450	1,554	0.891	0.623	2,891	0.264	0.04	7	270	5	185	7	41	10
MORB		unpb	AG-32-3-36	S. Atlantic	Bouvet	98	89.95	282	75474	1217	294027	1869	2239			427	1,612	0.761	0.575	2,476	0.144	0.04	4	264	4	156	4	40	6
MORB		unpb	AG-32-3-41	S. Atlantic	Bouvet	73	90.22	301	74107	1199	297481	1902	2432			474	1,618	0.818	0.606	2,566	0.132	0.02	5	149	3	122	10	35	7
MORB		(S7)	518-60/1	S. Atlantic	Bouvet	5	86.70	225	99549	1199	282320	1115	3478			536	1,204	1.232	0.759	1,120	0.99	0.03	5	308	6	345	11	24	22
MORB		unpb	13-12/49a	Atlantic Ocean	Romanche F.Z.	60	88.35	198	80862	1332	295314	2332	1999	139	270	1,648	0.677	0.547	2,968	0.070	0.05	3	230	4	206	6	13	1	6
MORB		unpb	13-12/49c	Atlantic Ocean	Romanche F.Z.	58	88.96	198	82768	1357	299084	2495	1953	145	264	1,640	0.672	0.556	2,966	0.087	0.04	5	288	5	170	13	13	1	5
MORB		unpb	13-12/49d	Atlantic Ocean	Romanche F.Z.	19	88.90	185	83795	1388	291945	2415	1872	143	250	1,656	0.641	0.537	2,881	0.063	0.06	7	415	7	229	31	18	2	10
MORB		unpb	13-12/49e	Atlantic Ocean	Romanche F.Z.	36	87.69	185	92539	1514	286753	1715	2076	154	322	1,636	0.724	0.670	2,854	0.095	0.02	4	102	3	160	4	10	2	5
MORB		unpb	13-12/49f	Atlantic Ocean	Romanche F.Z.	17	88.39	195	80132	1331	293645	2443	1985	141	260	1,661	0.676	0.542	3,048	0.044	0.06	6	478	9	339	11	12	2	8
MORB		unpb	13-12/49g	Atlantic Ocean	Romanche F.Z.	21	88.23	201	81299	1338	292863	2402	1996	145	262	1,646	0.681	0.554	2,954	0.074	0.06	3	438	8	358	11	15	2	6
MORB		unpb	13-11/1	Atlantic Ocean	Romanche F.Z.	21	88.11	240	89219	1358	287667	1705	1996	154	307	1,522	0.694	0.519	1,911	0.330	0.01	3	75	2	130	2	9	2	4
MORB		unpb	13-12/78	Atlantic Ocean	Romanche F.Z.	36	87.44	205	94900	1336	285163	1854	1568	164	341	1,420	0.550	0.617	1,970	0.542	0.05	3	341	6	219	6	11	4	
MORB		unpb	13-12/66	Atlantic Ocean	Romanche F.Z.	2	90.56	283	70902	1193	296004	2433	2279	130	356	1,682	0.770	0.546	3,432	-0.001	0.01	13	253	0	1571	18	8	4	
MORB		unpb	13-12/78	Atlantic Ocean	Romanche F.Z.	13	88.34	211	79948	1314	291469	2395	2059	147	283	1,643	0.707	0.565	2,996	0.080	0.08	5	562	10	433	18	23	3	
MORB		unpb	13-13/5	Atlantic Ocean	Romanche F.Z.	23	88.16	230	81451	1309	291485	1722	1767	161	357	1,607	0.696	0.494	2,114	0.154	0.03	2	238	3	195	3	2	2	
MORB		unpb	13-13/14	Atlantic Ocean	Romanche F.Z.	2	88.27	164	87742	1448	287388	1883	1501	142	284	1,651	0.522	0.498	2,146	0.064	0.08	0	653	0	39	11	8	0	
MORB		unpb	22-40	Arctic ocean	Knipovich ridge	31	87.56	213	93547	1512	286542	1813	1775	155	288	1,616	0.619	0.579	1,938	0.136	0.04	6	216	4	350	4	35	8	
MORB		(S6)	22-23	Arctic ocean	Knip																								

GROUP	Geodynamic setting	References	Sample	Geographic	Locality	N	Fe	Al ppm	Fe ppm	Mg ppm	Ca ppm	Co ppm	Cr ppm	100*Ni/Fe	100*Ni/Fe	100*Ni/Mg	100*Co/Fe	X px Mn/Fe	100Ca/Fe	1000Ca/Fe	Fe wt%	Al ppm	Fe ppm	Mn ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm	
WPM-THIN	OIB	(S19)	03-140	Iceland	Ksuteill	45	88.56	305	85943	1311	289225	2161	2359	167	321	1.527	0.816	0.700	2.517	0.320	0.03	9	223	4	186	13	17	1	6	
WPM-THIN	OIB	(S19)	03-161	Iceland	Ksuteill	50	88.48	296	86408	1326	288657	2157	2341	169	317	1.535	0.811	0.701	2.497	0.303	0.04	10	263	6	215	8	17	1	6	
WPM-THIN	OIB	(S19)	01-155	Iceland	Theshtareykir/Picrites	24	90.30	356	73313	1187	296901	2439	2469	144	584	1.586	0.838	0.615	3.327	0.182	0.06	19	430	9	356	31	33	2	21	
WPM-THIN	OIB	(S19)	01-57-4	Iceland	Theshtareykir/Picrites	32	90.06	379	74874	1187	295369	2371	2467	589	1.586	0.835	0.625	3.167	0.199	0.04	15	268	6	343	22	26	1	15		
WPM-THIN	OIB	(S19)	01-56-1	Iceland	Theshtareykir/Picrites	48	90.21	299	73806	1189	295885	2435	2423	552	1.611	0.819	0.604	3.299	0.147	0.04	10	284	6	227	15	18	1	13		
WPM-THIN	OIB	(S19)	01-41	Iceland	Theshtareykir	15	89.76	437	77047	1178	294013	2177	2800	147	631	1.529	0.952	0.718	2.826	0.216	0.08	17	584	9	406	8	20	2	15	
WPM-THIN	OIB	(S19)	01-56-2	Iceland	Theshtareykir/Picrites	9	90.49	350	71973	1149	297638	2425	2491	153	586	1.586	0.836	0.602	3.270	0.177	0.10	18	480	17	608	76	73	4	53	
WPM-THIN	OIB	(S19)	01-54	Iceland	Theshtareykir/Picrites	45	87.35	293	90335	1407	287024	2183	2297	172	399	1.557	0.800	0.723	2.417	0.258	0.04	9	298	5	226	13	14	1	8	
WPM-THIN	OIB	(S19)	01-51	Iceland	Theshtareykir	6	91.39	367	65650	1035	303239	2433	2287	138	885	1.577	0.754	0.495	3.707	0.217	0.15	38	1176	25	359	108	107	4	71	
WPM-THIN	OIB	(S20)	03-224	Iceland	Snaefellsness	4	89.14	292	82132	1272	283273	1994	2363	150	513	1.549	0.806	0.662	2.428	0.275	0.25	12	1823	29	1054	17	86	3	27	
WPM-THIN	OIB	(S19)	03-226	Iceland	Snaefellsness	15	88.28	232	88383	1409	289892	2079	1984	161	364	1.594	0.685	0.605	2.352	0.181	0.08	9	580	11	736	69	33	3	16	
WPM-THIN	OIB	(S19)	03-220	Iceland	Snaefellsness	5	89.93	245	83462	1449	291920	2073	2067	156	458	1.611	0.708	0.591	2.483	0.251	0.21	23	1484	17	1033	93	55	6	32	
WPM-THIN	OIB	(S20)	SM-6	Azores	Sao Miguel	26	87.43	166	94610	1469	286286	2062	1706	161	308	1.552	0.596	0.564	2.180	0.268	0.05	3	341	5	243	16	19	2	5	
WPM-THIN	OIB	(S20)	SM-10	Azores	Sao Miguel	117	88.58	184	85252	1181	237785	1919	2387	151	553	1.385	0.829	0.707	2.251	0.615	0.01	2	78	2	79	7	6	1	3	
WPM-THIN	OIB	(S20)	SM-16	Azores	Sao Miguel	122	88.18	233	81082	1094	290782	1812	2312	149	672	1.349	0.795	0.645	2.234	0.689	0.02	2	148	3	100	10	26	1	3	
WPM-THIN	OIB	(S20)	SM-36a	Azores	Sao Miguel	13	88.88	222	75771	1140	282770	1922	2605	139	525	1.504	0.890	0.674	2.537	0.368	0.10	9	761	12	481	26	40	3	13	
WPM-THIN	OIB	(S20)	SM-36	Azores	Sao Miguel	75	87.91	170	89944	1397	284715	2357	1774	154	355	1.553	0.623	0.560	2.620	0.267	0.03	2	205	4	126	13	8	1	3	
WPM-THIN	OIB	(S20)	SM-4	Azores	Sao Miguel	4	88.20	177	87865	1297	265944	1867	1925	145	411	1.476	0.674	0.592	2.147	0.425	0.11	3	796	5	465	5	23	5	16	
WPM-THIN	OIB	(S20)	SM-33	Azores	Sao Miguel	59	88.68	163	84172	1289	287065	2394	2159	149	414	1.508	0.752	0.633	2.844	0.359	0.02	2	175	3	121	17	14	1	3	
WPM-THIN	OIB	(S20)	SM-8	Azores	Sao Miguel	3	88.45	199	86050	1203	286761	1894	2449	152	447	1.398	0.854	0.735	2.201	0.489	0.12	9	838	32	653	98	219	6	44	
WPM-THIN	OIB	unpb	T-18	Azores	Tercena	11	87.83	435	91782	1322	289114	1982	2489	552	1.441	0.854	0.783	2.159	0.439	0.04	17	904	14	376	8	28	16	16		
WPM-THIN	OIB	(S22)	BH-15	Atlantic Ocean	Atlantic Ocean	33	88.24	404	89264	1333	291557	2119	2386	510	1.483	0.818	0.730	2.374	0.391	0.14	7	326	5	230	3	17	7	7	7	
WPM-THIN	OIB	(S22)	BH-19	Atlantic Ocean	Atlantic Ocean	16	87.83	483	92074	1353	289015	2151	2562	567	1.469	0.887	0.816	2.338	0.426	0.02	7	173	3	182	8	16	1	7	7	
WPM-THIN	OIB	(S22)	BH-34	Atlantic Ocean	Atlantic Ocean	24	87.77	410	92639	1335	289280	2043	2655	476	1.441	0.918	0.850	2.205	0.499	0.05	10	396	6	259	5	22	13	13	13	
WPM-THIN	OIB	(S22)	BOH-14	Atlantic Ocean	Atlantic Ocean	10	86.50	461	101517	1375	282492	1904	2701	411	1.355	0.965	0.969	1.875	0.677	0.12	19	875	12	594	9	32	12	12	12	
WPM-THIN	OIB	(S22)	BOH-24	Atlantic Ocean	Atlantic Ocean	14	89.35	437	90885	1195	294511	2043	2912	654	1.482	0.989	0.798	2.533	0.414	0.09	6	688	11	494	16	28	14	14	14	
WPM-THIN	OIB	(S22)	BOH-27	Atlantic Ocean	Atlantic Ocean	24	88.90	426	84083	1323	292930	2042	2849	626	1.466	0.973	0.818	2.428	0.446	0.06	8	423	6	267	8	8	7	7	7	
WPM-THIN	OIB	(S22)	BOH-33	Atlantic Ocean	Atlantic Ocean	20	88.38	419	87922	1282	293959	1956	2782	550	1.435	0.956	0.841	2.227	0.510	0.14	7	305	7	295	7	16	11	11	11	
WPM-THIN	OIB	(S22)	AM-7a	Atlantic Ocean	Atlantic Ocean	18	86.21	407	102948	1438	280159	1956	2396	328	1.397	0.855	0.880	1.900	0.580	0.06	13	430	7	303	20	18	11	11	11	
WPM-THIN	OIB	(S22)	BB-22	Atlantic Ocean	Atlantic Ocean	12	86.68	319	98731	1345	282319	1621	2549	385	1.349	0.903	0.900	1.626	0.690	0.12	53	785	19	654	121	63	32	32	32	
WPM-THIN	OIB	(S22)	B-26	Atlantic Ocean	Atlantic Ocean	3	86.66	450	99990	1417	282539	2101	2606	481	1.417	0.922	0.922	2.101	0.547	0.31	52	2177	39	1420	33	97	75	75	75	
WPM-THIN	OIB	(S22)	B-29	Atlantic Ocean	Atlantic Ocean	12	87.29	393	95617	1355	285940	1955	2696	448	1.417	0.943	0.902	2.044	0.547	0.09	21	613	10	532	13	23	15	15	15	
WPM-THIN	OIB	(S23)	BB	Atlantic Ocean	Atlantic Ocean	1	92.65	386	55709	898	305305	2051	3450	118	828	1.613	1.129	0.629	3.882	0.143	0.00	14	564	6	381	12	9	2	14	14
WPM-THIN	OIB	(S24)	95976	Atlantic Ocean	Atlantic Ocean	16	92.05	487	60473	944	304500	1894	3749	125	886	1.562	1.231	0.745	3.132	0.249	0.08	14	390	4	286	11	14	1	16	16
WPM-THIN	OIB	(S24)	27141	Atlantic Ocean	Atlantic Ocean	32	90.93	469	68080	1032	268916	1944	3499	136	848	1.515	1.178	0.802	2.866	0.344	0.05	14	390	7	277	16	24	2	12	12
WPM-THIN	OIB	(S24)	G-27142	Atlantic Ocean	Greenland	21	91.06	379	67782	1062	300460	1372	3128	143	725	1.567	1.041	0.706	2.762	0.236	0.07	8	500	7	277	16	24	2	12	12
WPM-THIN	OIB	unpb	Dis-1	Atlantic Ocean	Disko	75	88.50	310	86270	1329	288874	2028	3006	468	1.541	1.040	0.887	2.350	0.292	0.03	6	217	3	142	8	6	6	6	6	
WPM-THIN	OIB	(S10)	1185A-9R-2-92-96	Pacific Ocean	Ontong Java Plateau	80	87.36	250	94242	1463	283514	2121	2156	167	377	1.553	0.760	0.717	2.250	0.267	0.02	3	157	3	124	6	7	1	3	3
WPM-THIN	OIB	(S10)	1187A-10R-7-46-49	Pacific Ocean	Ontong Java Plateau	15	86.76	269	83907	1115	289525	1584	3313	152	743	1.328	1.149	0.964	1.897	0.731	0.08	6	604	8	362	15	48	2	30	30
WPM-THIN	OIB	(S10)	1187A-6R-5-116-119	Pacific Ocean	Ontong Java Plateau	4	88.33	292	79928	1202	291333	1910	2448	157	506	1.504	0.840	0.672	2.390	0.367	0.07	25	377	7	704	32	70	3	41	41
WPM-THIN	OIB	(S25)	P23-9	Africa	Karoo	21	88.28	333	87816	1245	287906	1623	2681	155	836	1.417	0.931	0.818	1.848	0.527	0.04	8	491	7	339	11	44	2	31	31
WPM-THIN	OIB	(S25)	P23-32	Africa	Karoo	25	88.49	361	79080	1138	293022	1632	2822	149	1106	1.439	0.963	0.761	2.063	0.484	0.05	7	303	4	240	9	11	2	24	24
WPM-THIN	OIB	(S25)	N356	Africa	Karoo	29	90.42	124	72082	922	296170	1094	4392	486	1.279	1.463	1.069	1.504	0.835	0.05	4	390	5	276	9	9	11	11	11	
WPM-THIN	OIB	(S26)	EM-43	China	Emnashan	26	90.47	256	72181	1286	298304	2612	2566	155	470	1.781	0.860	0.621	3.619	-0.207	0.04	14	284	5	301	39	18	1	24	24
WPM-THIN	OIB	(S27)	EM-55	China	Emnashan	5	90.59	213	71310	1036	298300	2084	3418	557	1.453	1.144	0.816	2.922	0.473	0.26	5	1869	29	1261	163	80	21	24	24	
WPM-THIN	OIB	(S27)	EM-57	China	Emnashan	4	90.69	198	71013	1020	301068	1728	3399	487	1.437	1.129	0.802	2.433	0.307	0.40	27	3014	41	1492	38	199	51	51	51	
WPM-THIN	OIB	(S27)	EM-58	China	Emnashan	4	90.76	208	69952	1001																				





GROUP	Geodynamic setting	References	Sample	Geographic	Locality	N	Fe	Al ppm	Fe ppm	Min ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm	100*Ni/Fe	100*Mn/Fe	100*Ni/Mg	100*Co/Fe	100*Cr/Fe	X ppm	Mn/Fe	100Ca/Fe	100Co/Fe	100Cr/Fe	Fe wt%	Al ppm	Fe ppm	Mn ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm
WPA/THICK	OIB	(S30)	SR-389-1.8-2.0	Hawaii	Mauna Kea, HSDP2	35	88.76	303	77207	1041	294410	15105	3425	147	795	1.349	1.163	0.898	1.949	0.680	0.04	2	297	4	232	3	42	2	11					
WPA/THICK	OIB	(S30)	SR-565-0.7	Hawaii	Mauna Kea, HSDP2	4	90.28	287	73350	962	296380	1261	4117	149	547	1.312	1.389	1.019	1.720	0.766	0.01	8	205	5	610	12	6	3						
WPA/THICK	OIB	(S30)	SR-682.9.8-1.0	Hawaii	Mauna Kea, HSDP2	50	88.70	289	84980	1138	298380	1576	3212	158	723	1.343	1.110	0.940	1.861	0.701	0.03	4	242	4	163	6	17	1	5					
WPA/THICK	OIB	(S30)	SR-694.5-0.5-5	Hawaii	Mauna Kea, HSDP2	28	88.91	284	83330	1111	290795	1550	3209	161	731	1.333	1.103	0.920	1.860	0.723	0.03	4	274	4	184	6	31	2	8					
WPA/THICK	OIB	(S30)	SR-690.4-0	Hawaii	Mauna Kea, HSDP2	16	88.67	307	77889	1032	294148	1587	3651	150	723	1.325	1.241	0.967	2.038	0.738	0.08	4	622	5	293	10	18	3	10					
WPA/THICK	OIB	(S30)	SR-694-3.0-3.2	Hawaii	Mauna Kea, HSDP2	2	89.01	278	82822	1104	292028	1601	3544	157	770	1.332	1.214	1.005	1.933	0.723	0.07	13	428	4	525	7	63	6	65					
WPA/THICK	OIB	(S30)	SR-930.9-1	Hawaii	Mauna Kea, HSDP2	10	88.24	229	80768	1122	292698	1545	3395	155	695	1.389	1.160	0.937	1.913	0.605	0.10	9	746	10	575	27	85	3	27					
WPA/THICK	OIB	(S30)	SR-942.5-7	Hawaii	Mauna Kea, HSDP2	40	88.18	276	80768	1122	292698	1545	3395	155	695	1.389	1.160	0.937	1.913	0.605	0.10	9	746	10	575	27	85	3	27					
WPA/THICK	OIB	(S30)	SR-954-10-7	Hawaii	Mauna Kea, HSDP2	48	88.92	289	79227	1099	290821	1487	3631	147	779	1.353	1.249	0.977	1.900	0.680	0.03	4	212	3	153	2	13	1	5					
WPA/THICK	OIB	(S30)	SR-961-14.0	Hawaii	Mauna Kea, HSDP2	44	88.44	261	79445	1082	292711	1506	3537	147	765	1.362	1.220	0.960	1.895	0.662	0.04	4	311	4	195	3	29	7						
WPA/THICK	OIB	(S30)	SR-962-18.1	Hawaii	Mauna Kea, HSDP2	66	88.51	284	79029	1075	293375	1488	3580	152	773	1.361	1.208	0.960	1.892	0.665	0.03	4	208	3	147	3	15	1	5					
WPA/THICK	OIB	(S31)	KI-11	Hawaii	Mauna Kea post shield	18	86.98	132	98166	1423	295490	2424	2233	366	1450	0.782	0.782	0.768	2.470	0.480	0.07	6	472	8	309	20	14	9						
WPA/THICK	OIB	(S31)	KI-8	Hawaii	Mauna Kea post shield	9	88.26	181	89202	1229	291934	1832	2448	536	1378	0.839	0.748	0.830	1.378	0.330	0.12	8	718	14	604	30	60	10						
WPA/THICK	OIB	(S31)	KI-4	Hawaii	Mauna Kea post shield	11	87.74	139	92943	1295	289526	2043	2588	446	1393	0.894	0.831	2.058	0.588	0.11	9	891	11	499	68	30	18							
WPA/THICK	OIB	(S31)	KI-3	Hawaii	Mauna Kea post shield	8	88.79	251	84979	1150	292785	1590	2916	654	1353	0.996	0.846	0.846	1.871	0.680	0.12	11	901	17	595	29	100	13						
WPA/THICK	OIB	(S31)	KI-3(n)	Hawaii	Mauna Kea post shield	5	88.82	234	84025	1145	290374	1562	2951	159	642	1.362	1.016	0.854	1.859	0.661	0.21	11	1510	28	1002	68	134	3	20					
WPA/THICK	OIB	(S32)	Koo-10	Hawaii	Koolau, Maekapu	7	87.13	179	96799	1263	285148	1274	2967	479	1305	1.040	1.007	1.316	0.780	0.14	7	1030	10	495	8	24	8							
WPA/THICK	OIB	(S32)	Koo-10a	Hawaii	Koolau, Maekapu	19	88.11	241	82175	1085	292709	1199	3910	613	1321	1.336	1.098	1.459	0.747	0.08	6	541	11	417	13	98	12							
WPA/THICK	OIB	(S32)	Koo-10b	Hawaii	Koolau, Maekapu	43	88.82	248	84276	1111	291260	1257	3707	610	1318	1.273	1.073	1.492	0.753	0.04	4	275	5	170	8	36	6							
WPA/THICK	OIB	(S32)	Koo-10c	Hawaii	Koolau, Maekapu	2	86.74	180	96604	1348	283591	1354	3505	404	1353	1.236	1.231	1.360	0.881	0.04	11	489	8	362	32	47	21							
WPA/THICK	OIB	(S32)	Koo-49	Hawaii	Koolau, Maekapu	34	88.00	235	83005	1056	292194	1272	3803	579	1320	1.302	1.080	1.532	0.748	0.03	5	227	6	212	7	37	10							
WPA/THICK	OIB	(S32)	Koo-55	Hawaii	Koolau, Maekapu	41	88.52	215	87012	1156	292015	1294	3584	557	1329	1.227	1.068	1.488	0.731	0.03	5	213	4	165	6	18	7							
WPA/THICK	OIB	unpb	S10(i)	Hawaii	Koolau, Maekapu	23	86.78	194	98745	1289	281959	1239	3129	171	475	1.306	1.110	1.096	1.254	0.751	0.04	3	307	5	213	4	165	6						
WPA/THICK	OIB	unpb	S10(i)	Hawaii	Koolau, Maekapu	2	87.86	230	90851	1142	286895	1194	4503	169	482	1.257	1.574	1.430	1.314	0.879	0.35	3	2433	12	1667	21	149	5	24					
WPA/THICK	OIB	(S33)	R8-1.1-1197.1	Hawaii	Koolau, KSDP	23	88.67	248	85157	1136	289945	1366	3516	152	620	1.334	1.213	1.033	1.604	0.720	0.04	6	299	5	323	14	49	2	7					
WPA/THICK	OIB	(S34)	LO-02-04	Hawaii	Loihi smt	28	87.70	236	91993	1283	285314	1984	2753	528	1394	0.965	0.888	2.156	0.595	0.05	7	395	7	248	22	29	10							
WPA/THICK	OIB	(S34)	LO-02-02	Hawaii	Loihi smt	19	88.29	219	88152	1233	289803	2036	2811	542	1389	0.972	0.857	2.309	0.585	0.07	5	515	8	282	66	35	17							
WPA/THICK	OIB	(S35)	158-9	Hawaii	Loihi smt	44	88.15	239	81977	1166	292971	1781	3100	659	1423	1.058	0.867	2.172	0.536	0.03	5	193	3	131	15	12	6							
WPA/THICK	OIB	(S35)	168-5	Hawaii	Loihi smt	3	88.43	309	79637	1131	283236	2015	2879	712	1420	0.981	0.981	0.782	2.531	0.542	0.34	14	2415	45	1594	84	123	51						
WPA/THICK	OIB	(S35)	187-1	Hawaii	Loihi smt	7	89.67	303	78190	1095	295195	1774	3242	679	1401	1.098	0.859	2.269	0.582	0.12	16	888	14	488	36	121	29							
WPA/THICK	OIB	(S36)	M2343-7	Hawaii	Loihi smt	9	88.14	215	89237	1319	288692	2194	2742	445	1478	0.950	0.848	0.878	2.298	0.576	0.08	6	585	12	312	64	40	10						
WPA/THICK	OIB	(S36)	M2335-12	Hawaii	Loihi smt	13	88.44	258	79807	1143	294008	1897	3091	683	1432	1.051	0.839	2.365	0.516	0.08	11	619	11	378	29	28	23							
WPA/THICK	OIB	(S36)	M2340-3	Hawaii	Loihi smt	3	87.56	222	93624	1291	286763	1911	3023	511	1379	1.054	0.967	2.041	0.628	0.32	26	2393	37	1203	88	156	39							
WPA/THICK	OIB	(S36)	M2343-10	Hawaii	Loihi smt	37	87.16	210	95978	1382	283538	2272	2781	550	1440	0.981	0.941	0.941	2.367	0.500	0.03	5	186	4	157	11	8	8						
WPA/THICK	OIB	(S36)	M2343-8	Hawaii	Loihi smt	5	88.16	210	88961	1295	288238	2180	2804	434	1456	0.973	0.865	2.450	0.468	0.13	37	914	11	569	51	9	36							
WPA/THICK	OIB	(S36)	M2340-1	Hawaii	Loihi smt	3	88.21	279	88346	1221	287678	1899	2982	620	1382	1.030	0.910	0.910	2.149	0.620	0.17	40	1340	14	269	43	99	28						
WPA/THICK	OIB	(S36)	MK2337-4	Hawaii	Loihi smt	19	87.11	224	96131	1351	282756	1826	2823	468	1405	0.998	0.960	1.900	0.573	0.07	10	524	10	364	43	18	11							
WPA/THICK	OIB	(S36)	M2343-9	Hawaii	Loihi smt	17	87.09	228	96726	1358	283961	2223	2577	445	1404	0.908	0.878	2.298	0.576	0.08	6	585	12	312	64	40	10							
WPA/THICK	OIB	(S36)	M2343-5	Hawaii	Loihi smt	3	88.87	217	83863	1242	291338	2211	2792	504	1481	0.958	0.804	2.536	0.416	0.30	19	2098	23	1585	74	36	30							
WPA/THICK	OIB	(S36)	M2336-5	Hawaii	Loihi smt	10	86.86	220	98509	1385	283269	2150	2695	385	1406	0.951	0.937	2.183	0.570	0.10	9	734	13	489	44	56	14							
WPA/THICK	OIB	(S36)	M2336-4	Hawaii	Loihi smt	21	87.57	235	93496	1298	286500	1996	2714	534	1388	0.947	0.886	2.135	0.608	0.06	5	425	9	304	27	20	14							
WPA/THICK	OIB	(S36)	M2338-5	Hawaii	Loihi smt	5	88.54	231	86290	1221	290233	2311	3149	423	1414	1.085	0.936	2.679	0.533	0.19	12	1356	27	891	22	67	4							
WPA/THICK	OIB	(S																																

GROUP	Geodynamic setting	References	Sample	Geographic	Locality	N	Fo	Al ppm	Fe ppm	Min ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm	100*Cr/Fe	100*Ni/Fe	100*Ni/Mg	100*Ca/Fe	X px Mn/Fe	STE	Fo wt%	Al ppm	Fe ppm	Mn ppm	Mg ppm	Ca ppm	Ni ppm	Co ppm	Cr ppm
WPM-THICK	OIB	(S38)	953C-93R-06 45-55	Atlantic Ocean	Gran Canaria	4	82.31	167	128922	1748	261052	1930	2485	187	219	1356	0.952	1.227	1.497	0.674	0.20	12	1377	18	908	61	74	5	6	
WPM-THICK	OIB	(S38)	953C-93R-4 18-24	Atlantic Ocean	Gran Canaria	5	87.75	181	91591	1300	285495	1602	3239	168	416	1419	1.135	1.039	1.749	0.544	0.18	13	1284	16	702	53	180	3	26	
KOMATITES	KOMATITES	(S39)	MUN-24	Canada	Munro 2.7 Ga	65	92.34	382	58523	925	307010	1562	3316	1441	1580	1.080	0.632	2.869	0.210	0.03	4	223	4	182	4	4	4	6		
KOMATITES	KOMATITES	(S39)	MUN-414	Canada	Munro 2.7 Ga	63	89.42	254	80218	1238	294877	1625	2942	732	1543	0.998	0.800	2.028	0.288	0.03	3	205	4	166	5	5	8			
KOMATITES	KOMATITES	(S39)	Ch-11a	Canada	Munro 2.7 Ga	32	88.82	299	84803	1339	293033	1645	3034	791	1579	1.035	0.878	1.940	0.212	0.04	5	327	7	214	8	6	6	16		
KOMATITES	KOMATITES	(S40)	M-626	Canada	Alexo 2.7 Ga	11	92.81	334	55132	862	309510	1490	3461	1299	1564	1.118	0.617	2.703	0.243	0.10	7	772	13	474	7	10	20			
KOMATITES	KOMATITES	(S40)	M-767	Canada	Alexo 2.7 Ga	8	93.53	390	49829	801	313247	1484	3481	1308	1607	1.111	0.554	2.978	0.155	0.14	14	1082	13	575	10	41	30			
KOMATITES	KOMATITES	(S41)	M-686	Canada	Alexo 2.7 Ga	33	94.30	421	43881	719	315979	1484	3353	1263	1638	1.061	0.466	3.381	0.091	0.05	9	404	5	274	5	6	20			
KOMATITES	KOMATITES	(S41)	MZ-4	Zimbabwe	Bellingwe 2.7 Ga	9	92.60	361	56373	885	307189	1494	3630	128	1327	1.571	1.182	0.666	2.651	0.230	0.11	10	796	13	477	6	8	2		
KOMATITES	KOMATITES	(S42)	G-21	Canada	Gilmour 1.9 Ga	15	92.03	359	60841	948	305902	1762	3351	1058	1558	1.095	0.667	2.897	0.256	0.08	8	626	9	327	11	18	25			
KOMATITES	KOMATITES	(S42)	G-18	Canada	Gilmour 1.9 Ga	2	91.83	352	62561	968	305804	1787	3304	1071	1547	1.080	0.676	2.856	0.278	0.36	3	2662	46	1468	14	35	38			
KOMATITES	KOMATITES	(S42)	G-15	Canada	Gilmour 1.9 Ga	63	88.74	222	86277	1281	292437	1796	2190	652	1503	0.749	0.639	2.106	0.371	0.04	3	263	4	156	2	3	5			
KOMATITES	KOMATITES	(S42)	G-8a	Canada	Gilmour 1.9 Ga	70	88.67	244	86637	1284	291770	1793	2617	675	1500	0.897	0.768	2.094	0.377	0.03	6	209	3	124	3	5	9			
KOMATITES	KOMATITES	(S43)	GOR-94-19	Gorgona	Gorgona 0.9 Ga	71	91.44	549	64607	1051	300452	2140	3593	920	1626	1.196	0.773	3.313	0.115	0.02	4	126	2	117	2	5	5			
KOMATITES	KOMATITES	(S43)	GOR94-35	Gorgona	Gorgona 0.9 Ga	13	92.58	533	56770	930	308331	2120	3447	1167	1638	1.118	0.635	3.734	0.090	0.08	10	613	11	396	10	16	41			
KOMATITES	KOMATITES	(S43)	GOR94-34	Gorgona	Gorgona 0.9 Ga	27	91.50	473	64764	1042	303202	2274	3247	1003	1609	1.071	0.694	3.511	0.150	0.05	12	402	7	256	12	12	15			
KOMATITES	KOMATITES	(S43)	GOR94-32	Gorgona	Gorgona 0.9 Ga	18	92.79	544	54632	907	305920	2121	3482	1195	1661	1.138	0.622	3.882	0.043	0.07	18	527	9	266	12	10	28			

**Notes for Table S2a.** Concentrations and standard errors of mean (STE) are shown for elements in ppm and for Fo in mol%.

N-number of averaged high-Mg olivines (see text for definition). Samples marked (i) and (ii) represent different compositional groups of olivine within the same sample. Samples marked (n) represent an additional independent population of olivines analyzed for the same sample. References include samples description, unpb- unpublished. \*- Ontong Java Plateau samples were arbitrary classified as WPM-THICK, but actually may belong to WPM-THIN at the time of formation.

## News

### Quick-Ref



We hope that you enjoy this section of Earth Pages. It gives one geologist's personal views about some of the major stories that appeared in leading journals, such as *Geology*, *Science* and *Nature*, in the last month. The items have been written and compiled by Steve Drury author of a number of Earth science books, including *Stepping Stones: The Making of our Home World* (Oxford University Press, 1999) and *Image Interpretation in Geology (3rd edition)* (Taylor & Francis, and Blackwell Science (USA), 2001).

If you have comments on this month's items you can contact the author at [s.a.drury@open.ac.uk](mailto:s.a.drury@open.ac.uk)

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## News for May 2007

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### Magmas from the mantle and recycled crust

May 2007

Two processes result in the Earth's mantle consuming rocks that formed oceanic and continental crust: subduction of oceanic lithosphere, and delamination from the base of thickened continental crust. In both cases the rocks involved are likely to be dominantly basaltic in composition. Those from the continental crust probably include mafic layered complexes, representing magma chambers in which intermediate magmas had resulted from fractional crystallization of basalt magma, and undifferentiated mafic igneous rocks underplated to the crust. Highly fractionated materials from the upper continental crust may also make their way into the mantle in the sedimentary cover to subducted ocean-floor basaltic crust.

New magmas originate in various ways as partial melts of ultramafic mantle rocks. Yet 4 billion years worth or more of consumed masses of crustal rocks must increasingly become involved in the chemistry of mantle melting by adding to its heterogeneity. Mantle heterogeneity is well-established from several lines of evidence provided by isotopic and trace-element analyses of modern basaltic lavas erupted in different tectonic settings. Yet, judging the influence and role of recycled crust have so far been plagued by data that are ambiguous. One outcome of previous research is that some ocean island basalt magmas formed by partial melting of peridotite whose chemistry had been transformed previously by other melts that had flowed through it (see Herzberg, C. 2007. Food for a volcanic diet. *Science*, v. **316**, p. 378-379).

A large team of geochemists, combining forces (and their data) from Russia, Germany, Australia, France, Taiwan, Eritrea, Britain, the USA, the Netherlands and Iceland, have sought to reduce the ambiguities by focussing on the chemistry of olivine phenocrysts found in basaltic lavas, rather than that of whole-rock samples (Sobolev, A.V. and 19 others 2007. The amount of recycled crust in sources of mantle-derived melts. *Science*, v. **316**, p. 412-417). Dominantly basaltic crustal masses in the mantle would melt to form silica-rich magmas. Passing through mantle peridotite, such melts would transform parts of the mantle to olivine-free pyroxenite. Magmas derived by partial melting of pyroxenite in the upper mantle would be basaltic, but enriched in silicon and nickel, and depleted in magnesium, calcium and manganese partly retained in the residual pyroxenes. Olivines crystallising first from basalt magmas carry a chemical signature of the parental melt composition, and thus the source.

The olivine approach by Sobolev and colleagues provides evidence for recycled crust in products of all kinds of basaltic magmatism, ranging from a contribution of 5% in ocean-floor basalts formed at ridges to about 20% in within-plate basalts. The contribution of mantle transformed to pyroxenite by chemical interaction with melt from foundered crustal masses ranges from 10% in mid-ocean ridge basalts to 100% in within-plate basalts formed below thick continental lithosphere. These include the largest volcanic outpourings in Earth's history, in the form of continental flood basalts. The largest of these, the Siberian Traps, accompanied the largest mass extinction of the Phanerozoic at the end of the Permian Period.

Earth-Page



## GEOCHEMISTRY

# Food for a Volcanic Diet

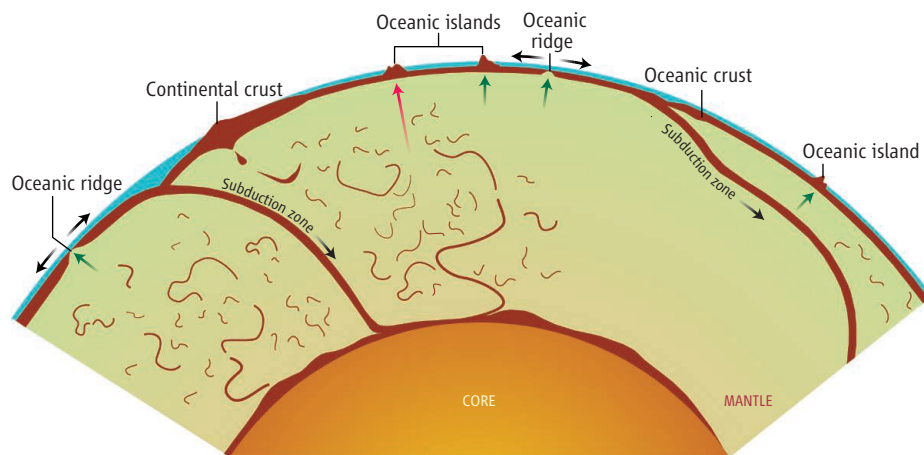
Claude Herzberg

Volcanic eruptions have the power to reshape Earth's landscape, alter climate, and affect life. To understand how this works requires that we go deep into the Earth to learn exactly what kind of rock melts to produce magmas and the chemistry of this source rock. These are fundamental problems in geology, and they are also among the most difficult to understand. On page 412 in this issue, Sobolev *et al.* (1) describe a method for identifying some of these source rocks. We can think of them as food for volcanoes in the sense that they melt to provide the magmas that can erupt to the surface. To understand what Sobolev *et al.* have done and the ramifications that go beyond Earth science, we need to start with a refresher in geology.

Earth's mantle consists mostly of peridotite, a rock rich in the mineral olivine  $(\text{Mg,Fe})_2\text{SiO}_4$ . When peridotite partially melts, the liquids collect to magmas that rise to the crust, give off gases like  $\text{SO}_2$ ,  $\text{CO}_2$ , and  $\text{H}_2\text{O}$ , and solidify to basalt, a rock rich in the minerals clinopyroxene  $[\text{Ca}(\text{Mg,Fe})\text{Si}_2\text{O}_6]$  and plagioclase  $[(\text{Ca,Na})(\text{Al}_{1-2}\text{Si}_{2-3})\text{O}_8]$ . Portions of these outer layers can be recycled back into the mantle at subduction zones and below thickened continents (see the figure). The recycled basaltic crust is transformed to a new rock called pyroxenite, so-called because it is rich in clinopyroxene. It may pile up on Earth's core, or be mixed back into the mantle with structures that have

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Chemical analyses of lava can now reveal the nature of the rocks deep in the Earth that melted and rose to generate specific volcanoes.



**Models of Earth's crust and mantle.** Oceanic crust (brown) is solidified liquid that forms by partial melting of mantle peridotite (green) at oceanic ridges; together with sediment, oceanic crust can be recycled back into the mantle at subduction zones (2, 3, 6). Continental crust (brown) forms at subduction zones and can be recycled when it thickens by delamination (5, 15). All crust (brown) is transformed to pyroxenite (brown) when recycled. Green arrow denotes melting peridotite. Red arrow denotes melting pyroxenite. Recycled crust may be distributed uniformly throughout the mantle, or it may be concentrated in certain hemispheres or depths. Crustal thickness is exaggerated for clarity, but ranges from ~6 to 40 km at the present time. Recycling is expected to reduce crust to dimensions ranging from micrometers to kilometers.

been described as marble cake (2), plum pudding (3), spaghetti (4), and gumbo (5) (see the figure). Volcanoes like those of Hawaii can melt from source rocks consisting of peridotite and/or pyroxenite from recycled crust. Sobolev *et al.* describe a method for identifying this rock based on the chemistry of lavas on volcanoes.

Sobolev *et al.* determined that many volcanoes melted from recycled crust, a conclusion that is not new (6). However, there has always been some ambiguity with past methods of identifying recycled crust based on the isotope

and trace-element geochemistry of lavas at the surface. New interpretations suggest that many oceanic islands melted from mantle peridotite that had been modified by melts that flowed through it (7, 8), a process called metasomatism (9). Because it makes no difference to an atom of lanthanum, for example, whether it ends up concentrated in the crust or as metasomatized peridotite, using it as a tracer can be ambiguous and nonunique (7, 8).

A breakthrough came when Sobolev *et al.* (10) showed that the nickel contents of many

olivine crystals in Hawaiian lavas were higher than those expected from melts of peridotite, and they preferred to explain this with a recycled crust source instead. But a lingering ambiguity is that a high nickel content in olivine can also arise when peridotite is enriched in pyroxene by melt-rock reaction (11). Supporting evidence for the recycled-crust interpretation (10) comes from the calcium contents of the Hawaiian lavas, which are too low to be easily explained by melting peridotite (12). Nickel and calcium are therefore telling the same story. That is, the main shield-building lavas at Hawaii were melted from a pyroxenite source rock that required the involvement of recycled crust as proposed by Sobolev *et al.* (10). The authors go further in that they examine the problem that arises when nickel, calcium, and manganese are used, and they extend the analysis to a larger population of volcanoes. Their results, together with other recent studies (7, 8, 11), show that it is unlikely that a single rock type will be an appropriate source for all oceanic volcanoes. For example, recycled

crust is an important source rock for the Hawaiian islands (1, 10, 12), whereas metasomatized peridotite is the source rock for the Cook islands (7). An outstanding question is whether peridotite sources become metasomatized by melted recycled crust (13) or in some other way (7, 8).

Future studies might allow us to transform our picture from hypothetical models to actual three-dimensional views showing the size and distribution of recycled crust in the mantle. The implications go far beyond geology. For example, it may be no surprise that Sobolev *et al.* (1) identify pyroxenite as the rock that melted to produce the Siberian Traps. This was a magmatic flood on land so massive in scale that it triggered the largest mass extinction of life on Earth, some 250 million years ago (14). Although the exact causal links remain poorly understood in detail, one can reasonably imagine a different outcome if the mantle diet had less pyroxenite and more peridotite. Under these circumstances, less magma would have been pro-

duced and made available for eruptive flooding, and Earth's biosphere could have evolved along different pathways.

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