Old subcontinental mantle zircon below Oahu

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Our understanding of mantle evolution suffers from a lack of age data for when the mantle geochemical variants (mantle components) developed. Traditionally, the components are ascribed to subduction of ocean floor over Earth history, but their isotopic signatures require prolonged storage to evolve. Here we report U-Pb age results for mantle-derived zircon from pyroxenite xenoliths in Oahu, Hawaii, using laser ablation inductively coupled plasma mass spectrometry and isotope dilution - thermal ionization mass spectrometry. The zircon grains have 14 million-year-old rims, Cretaceous cores, and Proterozoic Lu-Hf model ages which are difficult to reconcile with transport of the pyroxenites in the Hawaiian mantle plume because the ages would have been reset by high temperatures. We suggest the zircons may have been preserved in sub-continental lithospheric mantle. They possibly reached Oahu by asthenospheric transport after subduction at Papua New Guinea or may represent fragments of sub-continental lithospheric mantle stranded during Pangean breakup.

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The convecting mantle is isotopically heterogeneous, with at least four endmember “mantle components” identified in oceanic basalts¹, but component origins, ages, and distribution remain poorly understood. This compromises our understanding of mantle evolution. Three of the mantle components are present in oceanic island basalts (OIB; Fig. 1) and are commonly ascribed to subduction of ocean floor materials over Earth history^{1–3}. However, their radiogenic isotopic signatures appear impossible to generate and preserve if convective mantle homogenization occurs on time scales of ≤ 1 Ga (e.g.^{4–6}). Model ages^{7–9} and other considerations indicate that the EM1 and HIMU (Enriched Mantle 1 and High Mu = high U/Pb) mantle components are as old as Archean, EM2 appears Proterozoic^{7,8,10}, and all require storage in isolated reservoirs prior to return to the convecting, oceanic mantle. The fourth component is not expressed in OIB, is the source for MORB (Mid-Ocean Ridge Basalt) and appears to reflect melt extraction from the mantle over Earth history to form the crust. Oceanic Island Basalts show a spectrum of isotopic compositions between MORB and the OIB components (Fig. 1). Some rare xenoliths in OIB may represent the types of materials that melt to yield the range of isotopic compositions seen in OIB, including the endmember mantle component signatures. None have been more extensively studied than the Salt Lake Crater (SLC) garnet pyroxenite xenoliths (hereafter “pyroxenites”) containing high-pressure majoritic garnet, nanodiamonds and evidence for kimberlitic melt infiltration^{11–16}. We report ages for zircon recovered from SLC mantle xenoliths. They represent the first zircon ages for mantle rocks from a within-plate, central ocean tectonic setting and potentially have important implications for understanding the origin of chemical variability in the mantle. We propose that they support a testable, grand unifying hypothesis for chemical variation that involves Phanerozoic recycling of long-isolated, ancient subcontinental lithospheric mantle back into the convecting oceanic mantle.

Background

Garnet pyroxenite and spinel lherzolite xenoliths occur in Honolulu Volcanic Series (HV), post-shield-forming, ≤ 0.6 Ma,

Fig. 1 Plot of ϵ_{Nd} versus ϵ_{Hf} showing the fields for MORB, Hawaii (shaded), OIB, and the positions of endmember mantle component OIB; EM1, EM2 and HIMU. Also shown are averages for selected endmember component islands where Hf isotopic data were available (Go = Gough, Ke = Kerguelen, Ai = Aitutaki, Up = Upolu, SH = St. Helena, Ma = Marquesas, Tu = Tutilla). The Salt Lake Crater garnet pyroxenite minerals and Honolulu Volcanic rocks (HV) show depleted-mantle compositions and plot adjacent to the MORB field. Fields on the diagram (e.g. Hawaii), the review of HV isotopic compositions, and garnet pyroxenite isotopic ratios are from Bizimis et al.²¹. Averages for endmember mantle component islands were calculated using data downloaded from the GEOROC database.

basanitic to nephelinitic tuff at Salt Lake Crater, Oahu, Hawaii¹⁷. Oahu is ≤ 5 Ma old¹⁸ and built on ~ 80 – 85 Ma oceanic crust¹⁹. It occurs near the young, southern end of the Hawaiian island chain formed as Pacific ocean floor moved over a presumed mantle plume^{18,20}. Hawaiian volcanoes from Hawaii to Oahu lie along two geochemically distinct, geographically parallel lines referred to as the Kea and Loa trends²⁰; the pyroxenite xenoliths come from the Loa trend. SLC xenoliths have been studied extensively for five decades because they are the deepest-origin rocks yet found in the ocean basins. They can potentially shed light on mantle chemical variability, lithospheric recycling, and mantle convection; in short the evolution of the Earth^{12–16,21}. Geothermobarometry indicates that SLC pyroxenite xenoliths formed/last-equilibrated between 975 and 1300 °C, at pressures of ~ 10 – 35 kb (~ 35 – 100 km) in the deep oceanic lithosphere or upper asthenosphere^{14,15,21}. However, phase assemblages¹⁴, the presence of nanodiamonds¹³ and majoritic garnet¹¹ imply that they also experienced depths of 150–240 km with temperatures above 1300 °C.

The pyroxenites may be high-pressure igneous cumulates^{12,14,21–23} with their allotriomorphic granular textures reflecting recrystallization and re-equilibration similar to that documented in other mantle pyroxenites²⁴. Spinel lherzolites occur with the garnet pyroxenites in the SLC tuff. They give ≤ 2 Ga rhenium-depletion model ages²⁵ but the pyroxenites have zero-age Lu–Hf and Sm–Nd isotopic systematics²¹. The pyroxenite Hf–Nd–Sr isotopic compositions are “depleted”, but distinct from MORB/E-MORB and are similar to, though not identical to, the host HV rocks²¹ (Fig. 1). Both the pyroxenites and HV are unique within the ocean basins with high $^{176}\text{Hf}/^{177}\text{Hf}$ compared to $^{143}\text{Nd}/^{144}\text{Nd}$ ratios²¹. Thus, the pyroxenites could be Honolulu Volcanic magma chamber cumulates. The associated SLC spinel lherzolites have unradiogenic Os and radiogenic Hf signatures that indicate they are fragments of ancient, depleted MORB lithosphere²⁵. The composition of the HV may reflect melting of an ancient subducted oceanic slab package that includes basaltic crust (now pyroxenite), sediments, and underlying depleted-mantle lithosphere²¹. Complicating the interpretation of pyroxenite data is evidence for CO_2 ¹⁹ and H_2O ¹⁶ fluid metasomatism, as well as infiltration of kimberlite-like melts¹²; features the authors ascribed to Cenozoic processes in the Pacific oceanic mantle, but commonly associated with subcontinental lithospheric mantle (SCLM; e.g.²⁶). Age data for the pyroxenites can potentially reduce the number of viable formative hypotheses and inform on the role of SCLM in the production and preservation of mantle heterogeneity.

Results

We collected 38 garnet pyroxenite xenoliths, selected the 7 largest (18.6–138.8 g with $\leq 20\%$ attached tuff) for heavy mineral extraction, and recovered 7 zircon grains from 2 samples (Fig. S1). Sample and zircon descriptions appear in Supplementary Information 1. Zircon separation techniques given in the Methods detail that it is extremely unlikely the zircon grains are contaminants. Uranium–Pb dating was done by Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS) and solution Isotope Dilution Thermal Ionization Mass Spectrometry (ID-TIMS) with Hf isotopes by Multicollector ICPMS (MC ICPMS; see Methods). Garnet and clinopyroxene were multiply analyzed for trace- and minor elements by LA ICPMS and Electron Microprobe (EMP; see Methods with data in Supplementary Data 1).

A primitive mantle-normalized diagram (Fig. 2) shows that the zircon-bearing pyroxenites have garnet and clinopyroxene compositions very similar to those in other Oahu pyroxenite xenoliths reported in the literature²¹. Laser ablation U–Pb isotopic analyses were carried out on the 7 unpolished zircon grains and yielded

Fig. 2 Primitive mantle normalized diagrams for SLC pyroxenites. Diagram (a) is for clinopyroxene and (b) shows garnet from the SLC garnet pyroxenites including those that contained zircon (SLCX 12 and 47). Shaded fields show the range of clinopyroxene and garnet compositions reported for SLC garnet pyroxenites by Bizimis et al.²¹. Normalizing values from McDonough and Sun⁶² and element order from Sun and McDonough².

time-resolved profiles of $^{206}\text{Pb}/^{238}\text{U}$ ratios. Profiles from 3 grains indicate the presence of older cores. Rim and core portions were dated by *selectively* (constant isotopic signature) averaging the early and late parts of profiles. The late parts give the core ages assuming that there is no overlap of the beam with the rim but are otherwise minimum core ages. The 7 rim analyses scatter to the right of the Wetherhill concordia curve, probably due to biases in the small ^{207}Pb peak (Fig. 3). However, their $^{206}\text{Pb}/^{238}\text{U}$ ages overlap within error (Table 1) and give a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 13.9 ± 0.6 Ma (MSWD=0.97, Fig. 3). Deep profiles from three grains show a shift in Th/U ratios and are wide enough to obtain age information. These analyses of SLCX12 Gr2, Gr3 and SLCX47 Gr2 interior regions give ages of 45 ± 6 Ma, 70 ± 4 Ma and 101 ± 7 Ma (Fig. 3, Table 1). The interior ages may have minor mixing between the rims and older core components (e.g.²⁷) and we cannot be sure that the 15 μm deep laser pits intersected only one phase at a given depth. Three of the grains analyzed by LA-ICPMS were then dissolved and analyzed by ID-TIMS. These also gave diverse $^{206}\text{Pb}/^{238}\text{U}$ ages of 56.0 ± 0.8 , 61.9 ± 1.7 and 124.1 ± 0.2 Ma but are certainly “mixtures” that reflect young rims and older cores. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages are as old as 1992 ± 141 Ma but may be affected by small amounts of common

Pb that bias the ^{207}Pb isotope (Table 1; Fig. 3). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios give ϵ_{Hf} values of -8.4 ± 0.6 and -1.6 ± 4.1 (Table 1) which are well below literature values for SLC pyroxenite clinopyroxene and garnet as well as the Honolulu Volcanic rocks shown in Fig. 1. The negative ϵ_{Hf} values have corresponding minimum depleted-mantle model ages (depleted-mantle consistent with Hf, Nd and Sr isotope data in ref. ²¹) of 1439 ± 137 Ma and 1037 ± 276 Ma, respectively, assuming a protolith $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.013 characteristic of felsic crust (see “Methods” for calculation procedures). A higher ratio of 0.022, which is more reflective of mafic crust, would result in older model ages of about 2200 Ma and 1600 Ma, respectively.

Discussion

The zircon grains and their ages can potentially constrain various models for the origin of chemical variability in the mantle. However, these are the first zircon ages ever reported for mantle rocks from a within-plate, central ocean tectonic setting. We propose some potential implications but note that their full ramifications will become clearer if future experiments recover zircon from other mantle xenoliths in the central ocean basins.

Fig. 3 Wetherill Concordia diagram showing zircon U-Pb dating results.

See legend at the bottom for the types of data represented by ellipses on the plot. The bar graph inset shows the weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age for the seven zircon rim analyses. ID-TIMS = isotope dilution thermal ionization mass spectrometry; LA-ICP-MS = laser ablation inductively coupled plasma mass spectrometry.

We begin below by discussing the dating results and then assess hypotheses to explain the ages.

As noted in the results, the outside rims on the seven zircon grains yielded overlapping ages of ~ 14 Ma, with three interior zones in two grains giving minimum ages of ~ 45 , 70 and 100 Ma. Their TIMS $^{207}\text{Pb}/^{206}\text{Pb}$ ages are as old as ~ 2000 Ma and the two solution Hf analyses gave old model ages between ~ 1000 and 2200 Ma depending on assumptions. The old ages are important. The garnet and clinopyroxene analyses from the seven xenoliths examined for zircons (two yielded the seven zircon grains) have similar trace element patterns that overlap those reported by²¹ (Fig. 2). However, our zircon grains give $\sim 5 \text{ } \epsilon_{\text{Hf}}$ values (Table 1) which are distinctly unradiogenic compared to the $+10$ to $+20 \text{ } \epsilon_{\text{Hf}}$ values reported by²¹ for pyroxenite clinopyroxene and garnet (Fig. 3). Further, these two major minerals have zero-age Lu/Hf isochron systematics. The difference between the Hf isotopic composition of pyroxenite zircon versus garnet and clinopyroxene suggests the major minerals “recently” recrystallized, or melted and recrystallized, on a time scale short enough that Hf in zircon did not diffusively re-equilibrate with the rest of the rock. Thus, zircon Hf preserves evidence of an ancient past for the pyroxenites and this conclusion is supported by the old $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

Zircon occurs in subcontinental mantle xenoliths/xenocrysts from lamproites and kimberlites^{28,29} but the mantle-derived zircon grains reported here are the first from a within-plate, central ocean basin and the zircon age information is all that is available for rocks from this setting. Although the pyroxenites containing the zircons have geochemical similarities to host Honolulu Volcanic (HV) rocks (e.g. high $^{176}\text{Hf}/^{177}\text{Hf}$ compared to $^{143}\text{Nd}/^{144}\text{Nd}$ ratios²¹), they cannot be HV cumulates; all the zircon ages, including the ~ 14 Ma rim ages, are older than the island of Oahu¹⁹. It is unlikely they are igneous cumulates associated with formation of the MORB crust underlying Oahu because pyroxenite Hf-Nd-Sr isotopic compositions are different from MORB/E-MORB²¹ (Fig. 1). However, it is possible that the ~ 70 Ma interior zircon age reflects zircon growth in a pre-existing

block of rock in association with the formation of the 80 Ma ocean floor underlying Oahu. It would reflect the time of closure following “capture” of the block as the new oceanic lithosphere cooled and thickened. Unexplained is the ~ 100 Ma date. Despite zero-age Lu-Hf and Sm-Nd isotopic systematics²¹, our old Hf (~ 1000 – 1400 Ma) and ~ 2 Ga $^{207}\text{Pb}/^{206}\text{Pb}$ “model” ages resemble the <2 Ga model rhenium-depletion ages²⁵ for associated SLC spinel lherzolites. Given the old model ages, we examine possible relationships between the xenoliths, mantle chemical variability and the mantle components.

The isotopically-identified EM1, HIMU and EM2 (Enriched Mantle 2) endmember mantle components in OIB (Fig. 1) have long been related to continuous subduction recycling of ocean floor basalt and sediment back into the mantle^{1–3,30}. Some OIB, such as the HV, have somewhat “depleted” compositions (Fig. 1) that may reflect melting of mantle component materials intermixed with depleted MORB mantle^{1–3}. If HV magma formation involved melting of the SLC garnet pyroxenites (isotopic compositions largely overlap) our old zircon ages, and ancient model ages, imply that “mixing” occurred long before the melting that produced the young HV rocks. Thus, the xenoliths may shed light on the origin of chemical variation in the mantle.

The distinct radiogenic isotopic signatures for the endmember mantle components imply age significance and they may not have been continuously produced^{7–10}. Continuous production predicts that recently subducted materials will not have had time to develop radiogenic mantle component signatures. However, ratios of similarly incompatible elements are relatively unaffected by percentages of melting and reflect mantle source compositions. These data have not revealed any OIB with endmember mantle component trace element ratios that were not also identified using isotopes^{7,30}. Various numerical modeling experiments indicate that OIB mantle component signatures should be obliterated by homogenizing mantle convection within 1 Ga^{4–6}. Creation and preservation of the mantle component isotopic anomalies appear to require prolonged reservoir isolation. We examine three recent models for component formation, isolation, and the preservation of distinct isotopic signatures in light of the zircon dates.

One possibility is that subducted ocean floor materials (e.g. now garnet pyroxenites) are stored for extended periods in the lower mantle and brought up by deep mantle plumes^{31–35}. Seismic tomography studies show that Large Low Shear-Wave Velocity Provinces (LLSVPs) representing high temperature/low density areas in the lower-most mantle connect to “plumes” rising to beneath Large Igneous Provinces (LIPS) and OIB^{32,35}. Although large-scale, mantle component, whole-rock isotopic anomalies may survive for billions of years in a lower mantle reservoir, the preservation of isotopic clocks in individual minerals, even zircon, appears unlikely. Mantle temperatures in the LLSVP are at least ~ 3300 °C³⁶ with temperatures at the top of the Hawaiian plume >1300 °C³⁷. Lead diffusion rates³⁸ show that 100 μm zircon is an open system at 1300 °C. Our 30 to 50 μm grains preserve zones with various “ages” (e.g. ~ 13.9 , 40, 70, and 100 Ma). They plot close to Concordia, which suggests little Pb-loss. It follows that they could not have been stored in the LLSVP and brought up by a plume. This is consistent with a study of continental flood basalts (LIPS) showing that numerous lithophile similarly incompatible element ratios and siderophile trace element signatures, correlate with the age (Archean versus Proterozoic and younger) of lithosphere transected⁸. Thus, LIP magmas may be created by the thermal energy from plumes, but their trace element signatures, which resemble the OIB mantle components, are determined by the age and composition of the SCLM they partially melted. This would not be expected if LLSVP plumes were the carriers of the mantle component signatures. In

Table 1 Isotopic data and ages for Salt Lake Crater, Garnet Pyroxenite Xenolith Zircon.												
Sample/grain	U (ppm)	²⁰⁶ Pb (ppm)	Th/U	²⁰⁷ Pb/ ²³⁵ U	2σ	²⁰⁶ Pb/ ²³⁸ U	2σ	Err. Corr. ^a	²⁰⁷ Pb/ ²³⁵ U	2σ	²⁰⁶ Pb/ ²³⁸ U Age Ma	2σ
A: Laser ablation results												
Exterior												
SLCX-47 Gr1	137	0.3	0.42	0.0808	0.0252	0.00201	0.00025	0.39160	78.9	23.6	12.9	1.6
SLCX-47 Gr2 ^b	74	0.2	0.30	0.0190	0.0074	0.00245	0.00047	0.50172	19.1	7.3	15.7	3.1
SLCX-47 Gr3	177	0.4	0.41	0.0698	0.0209	0.00222	0.00021	0.32190	68.5	19.7	14.3	1.4
SLCX-12 Gr1 ^b	168	0.4	0.40	0.0958	0.0308	0.00218	0.00026	0.36878	92.9	28.3	14.0	1.7
SLCX-12 Gr2	611	1.4	0.22	0.0184	0.0035	0.00234	0.00030	0.68592	18.5	3.4	15.1	1.9
SLCX-12 Gr3 ^b	301	0.6	0.26	0.0135	0.0023	0.00213	0.00030	0.85465	13.6	2.3	13.7	2.0
SLCX-12 Gr4	500	1.0	0.18	0.0275	0.0071	0.00210	0.00016	0.29950	27.5	7.0	13.5	1.0
Interior												
SLCX-47 Gr2	313	4.9	0.39	0.1011	0.0207	0.01580	0.00110	0.34008	97.8	19.0	101.0	7.0
SLCX-12 Gr2	635	4.4	0.43	0.0447	0.0141	0.00700	0.00088	0.39654	44.4	13.7	45.0	5.6
SLCX-12 Gr3	914	9.9	0.34	0.0696	0.0077	0.01084	0.00055	0.46289	68.3	7.3	69.5	3.5
Sample/grain	Wt. (μg)	U ppm	Th/U ^c	Pb _T pg ^d	Pb _C pg ^e	²⁰⁶ Pb/ ²⁰⁴ Pb _m f	2σ	²⁰⁶ Pb/ ²³⁸ U ^g	2σ	Err. Corr. ^a	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ
B: U-Pb chemical abrasion ^f ID-TIMS isotopic and age data												
SLCX-47 Gr2 ^b	0.5	182	0.61	1.88	0.24	484.5	36.5	0.1203	0.0013	0.019440	0.000027	0.00046
SLCX-12 Gr3 ^b	1.0	23	0.42	0.23	0.29	67.1	18.5	0.0780	0.0233	0.009654	0.000262	0.01633
SLCX-12 Gr1 ^b	1.0	139	0.30	1.27	0.36	235.2	42.2	0.1473	0.0139	0.008727	0.000119	0.00988
B: U-Pb chemical abrasion ^f ID-TIMS isotopic and age data												
SLCX-47 Gr2 ¹	124.1	0.2	115.35	4.4	~	~	~	~	~	~	~	~
SLCX-12 Gr3 ¹	61.9	1.7	76.3	22.0	553.3	1991.7	616.0	89	89	140.6	98	98
SLCX-12 Gr1 ¹	56.0	0.8	139.5	12.3	1991.7	1991.7	140.6	98	98	140.6	98	98
¹⁷⁶ Hf/ ¹⁷⁷ Hf _m k	2σ	ε _{Hf} ¹	2σ	T _{DM} Ma ¹	2σ	T _{CHUR} Ma ¹	2σ	% Disc ¹	2σ	% Disc ¹	2σ	2σ
C: Hf isotopic data and model ages												
0.282542	0.000017	-8.4	0.6	1439	137	628	89	89	89	140.6	98	98
0.282734	0.000116	-1.6	4.1	1037	236	132	306	306	306	140.6	98	98
a Err. Corr. — error correlation coefficient for concordia coordinates												
b Grains used for TIMS isotopic analysis and dating												
c Th/U calculated from radiogenic ²⁰⁶ Pb/ ²⁰⁶ Pb ratio and ²⁰⁷ Pb/ ²⁰⁶ Pb age assuming concordance												
d Pb _T = total Pb in picograms excluding blank												
e Pb _C is total common Pb in picograms assuming the isotopic composition of laboratory blank assigned the composition ²⁰⁶ Pb/ ²⁰⁴ Pb = 18.49 ± 0.4%, ²⁰⁷ Pb/ ²⁰⁴ Pb = 15.59 ± 0.4%, ²⁰⁸ Pb/ ²⁰⁴ Pb = 39.36 ± 0.4%												
f ²⁰⁶ Pb/ ²⁰⁴ Pb _m and ²⁰⁷ Pb/ ²⁰⁴ Pb _m are measured values and corrected for fractionation and common Pb in the spike												
g Pb _T /U ratios corrected for fractionation, common Pb in the spike, and blank. ²⁰⁶ Pb/ ²³⁸ U and ²⁰⁷ Pb/ ²⁰⁶ Pb also corrected for ²³⁰ Th disequilibrium assuming Th/U = 4.2												
h Relation between ages and concordia coordinates: Y = ²⁰⁶ Pb/ ²³⁸ U = EXP(λ ₂₃₈ U (206Pb- ²³⁸ U Age)) - 1; X = ²⁰⁷ Pb/ ²³⁵ U = EXP(λ ₂₃₅ U (207Pb- ²³⁵ U Age)) - 1												
i See Methods for decay constants and ratios used in calculating ages and model ages as well as chemical abrasion methods												
j % Disc is percent discordance for the given ²⁰⁷ Pb/ ²⁰⁶ Pb age												
k ¹⁷⁶ Hf/ ¹⁷⁷ Hf _m = measured Hf isotopic ratio in zircon												
l See Methods for calculation of ε _{Hf} and Hf model ages = TDM (Depleted Mantle) and TCHUR (Chondritic Uniform Reservoir)												

summary, the LLSVP model provides a mechanism for the preservation of whole-rock isotopic signatures. However, it is unlikely that the radiometric ages of the zircons in the pyroxenites would survive long-term (>100 Ma) storage in the LLSVP before being entrained in a plume below Oahu.

To avoid homogenization by mantle convection, materials could also be stored in cool subcontinental lithospheric mantle^{1,6,10} (SCLM). The presence of significantly older zircon cores and ancient Hf model ages is consistent with zircon formation in a continental setting far from Oahu. Two models explain how materials might be returned to the convecting mantle: (1) subduction-related delamination of SCLM and (2) stranding of SCLM during rifting.

Regarding (1), it has been noted¹ (also see Anderson references therein), that the subduction assembly of Pangea may have delaminated ancient continental lithosphere creating the globe-encircling, oceanic “Dupal” anomaly bearing the mantle components. Radiogenic isotope studies (e.g.³⁹) have shown that even Archean SCLM can be delaminated. It has been proposed⁴⁰ that subduction zone processes, such as differentiation of arc magmas, and partial melting of subducting oceanic crust result in metasomatized SCLM, high-density cumulates and melt residues (arclogites) susceptible to delamination.

An object subducted and convectively delivered by the asthenosphere to Oahu requires a subduction zone that dips beneath the Pacific plate toward Oahu. The only part of the circum-Pacific system where this occurs is a ~5000 km long section from Papua New Guinea to the New Hebrides trench. Only the western section over Papua has subduction with volcanism occurring through continental crust. This crust includes late Proterozoic and Paleozoic basement of the Australian continent⁴¹ that is broadly similar in age to the cores, and the Hf model ages, found in the pyroxenite zircon. The implied transport direction is nearly normal to movement of the overlying Pacific plate. Lithospheric plate velocities and directions can be determined for the past 250 Ma but underlying asthenospheric flow velocities and trajectories are difficult to determine. Studies based on surface wave shear splitting^{42–44} suggest convective flow directions from the inferred orientation of mineral fabrics with relative rates of flow from the degree of anisotropy. Seismic anisotropy measurements in the asthenosphere beneath the Pacific plate⁴³ show that at shallow levels (ca. 50 km) fast directions east of Papua New Guinea are oriented NE and bend to E-W in the central Pacific. This pattern is broadly consistent with convective transport of subducted material from the Papua trench to the Hawaiian chain. Orientations tend to swing to the NW, the direction of plate motion, at deeper levels, but a central channel extending from the Papua region to the central Pacific shows little anisotropy⁴³. The inferred transport direction is most consistent with seismic anisotropy data if it occurred at shallow levels in the asthenosphere.

If for discussion it is assumed the Oahu pyroxenites originated from Papua New Guinea, the zircon data furnish estimates of asthenospheric flow rates. Oahu is ~6600 km ENE of the subduction zone. Assuming that arc-related rocks containing zircon were delaminated immediately after zircon crystallization allows ~14 Ma (age of zircon rims) for transport to beneath Oahu. Xenolith incorporation in the plume-related magma that generated the Salt Lake Crater volcanic rocks occurred < 1 Ma ago. This implies an average transport rate from convective flow in the shallow asthenosphere of ~50 cm/yr. If delamination at Papua New Guinea occurred long after crystallization, or if the zircons arrived prior to when the xenolith-bearing Honolulu volcanic rocks were produced, the rate would be higher.

The inferred convective flow rate is within a factor of 3 of present-day maximum ocean ridge spreading rates (15 cm/yr)⁴⁵ and within 2 of some past, Cenozoic rates such as for the Indian

Ridge (23 cm/yr)⁴⁶. Presumably plate movement rates are below those of mantle convection due to the energy losses from plate deformation, uplift and magmatism. It may be coincidental, but seems rather remarkable, that the distance to Papua New Guinea and the zircon rim ages just happen to yield a transport rate so similar to plate movement rates.

Viability of this hypothesis requires delamination of blocks of arc-affected (at ~14 Ma) subcontinental lithosphere thick enough to prevent interior heating to above the zircon closure temperature (<1000 °C) during transport in 1300 °C asthenosphere in ~14 M.Y. Recovery of several, same-age zircon grains in a single hand sample implies limited disaggregation of the delaminated material over this long distance. Numerical modeling indicates that lower continental lithosphere, weakened by melt infiltration above subduction zones, can result in the subduction-delamination of subcontinental lithospheric mantle⁴⁷. Thermal modeling equations⁴⁸ for cooling/warming by conduction indicates that a 100 km-thick SCLM slab, initially averaging 750 °C, and delaminated/subducted into 1300 °C asthenosphere, still has 60% of its volume below the ~1000 °C Pb closure temperature for 100 µm zircon³⁸ after 100 M.Y. (see Supplementary Information 2 for modeling details).

In summary, the zircon ages, subduction directions, and inferences on asthenospheric flow suggest the Oahu pyroxenites might have come from Papua New Guinea. Thermal modeling indicates that zircon could survive “resetting” if the delaminated lithosphere was reasonably thick. The hypothesis does not account for the ultra-high-pressure signatures in the Salt Lake Crater pyroxenite xenoliths (e.g. majoritic garnet, nanodiamonds) but, as discussed below, these characteristics may be relicts of earlier processes that can affect SCLM.

There is an alternative model involving rifting to explain the origin of the Oahu zircon grains and xenoliths. The present-day, overall E-W motion of tectonic plates dating back to the Mesozoic breakup of Pangea, suggests the globe-encircling Dupal anomaly¹ may be related to rifting. Neoproterozoic to Archean Re-depletion model ages have been reported for sulfide grains in mantle peridotite xenoliths from Sal Island, Cape Verde⁴⁹. The peridotites may contribute to the extreme HIMU to EM1 signatures seen in Cape Verde basalts. The model ages mirror the tectonic history of Western Africa and Eastern Brazil and the authors propose that the xenoliths are ancient, Archean, African/Brazilian SCLM left stranded in the convecting mantle during opening of the Atlantic Ocean. Plume-related stranding appears to have also left a sliver of continental crust, and perhaps subcontinental lithosphere, below southeast Iceland⁵⁰. Although the Cape Verde authors⁴⁹ report model ages, preservation of the isotopic information in small mineral grains implies that they withstood re-equilibration for billions of years and the subsequent, ~60 Ma separation of N. Africa from S. America. The authors proposed that the ages reflect storage in low geotherm, ancient SCLM that survived isotopic resetting because the removed SCLM did not experience prolonged, high, isotope-resetting, asthenospheric temperatures. The conductive modeling results (Fig. S2) suggest that zircon, in thick SCLM packages, can survive radiometric re-equilibration in the asthenosphere for times on the order of 100 Ma.

Various geochemical, mineralogical and geochronological characteristics of the Hawaiian pyroxenites place limits on their origin and perhaps support a rift hypothesis. It is conceivable that the zircons grew during deep addition of a kimberlite component at the margins of the spreading ridge that formed the ocean floor below Oahu ~80–85 M.Y. ago¹⁹. Most research on the pyroxenites has ascribed the nanodiamonds¹³, majoritic garnet¹¹ and kimberlite-like melt infiltration¹² to late Cenozoic processes that operated in the Pacific oceanic mantle. Individually, depleted mantle (as proposed for the preservation of the Cape Verde

xenoliths⁴⁹) plus kimberlite magma, diamonds, and majoritic garnets are rare characteristics of rocks from various geologic/tectonic environments, but together they are associated with cratonic lithospheric and sub-lithospheric mantle as demonstrated by numerous papers (e.g.^{26,51,52}). Although the pyroxenites examined here did not contain Archean zircon, their grains gave model ages suggesting a Paleoproterozoic history and zircon can form in SCLM long after cratonization as a result of mantle metasomatic processes²⁹. Thus, the Oahu zircon grains, their host pyroxenites, and the mantle components, may reflect SCLM stranded in the upper mantle during Phanerozoic rifting events that disaggregated Pangea. In this scenario, the 70 Ma zircon interior age could reflect heating of SCLM blocks when the crust below Oahu formed at the ridge some 80 Ma ago. They became “captured” as the lithosphere thickened. The younger 14 Ma zircon ages may somehow reflect later heating by the Hawaiian plume but if the pyroxenite block was attached to the NW-heading lithospheric plate, it is not obvious how it got heated 14 Ma before it was over the plume. The cratonic characteristics of the xenoliths and Cape Verde results⁴⁹ support a rifting model but our ~14 Ma zircon rim ages are better explained by the Papua New Guinea subduction model. These hypotheses can be tested by future experiments designed to identify xenolith zircon in endmember component OIB. With comprehensive dating, the geographic origin of xenoliths may be identifiable, and lead to a better understanding of how SCLM is returned to the convecting mantle to generate mantle geochemical heterogeneity.

Methods

Sampling and zircon separation. Thirty eight garnet pyroxenite xenoliths were collected from the basaltic to nephelinitic tuff at Salt Lake Crater Oahu, Hawaii within ~20 m of N 21° 21.174', W 157° 54.727'. Optical microscopy and Scanning Electron Microscope (SEM) backscatter images failed to reveal thin section zircons for in-situ dating. Zircons are rare in primitive mafic rocks (e.g. zircon-bearing SLCX 47 contains Mg-rich, Fo₇₉ olivine) and so the lack of grains in thin section was not unexpected.

The seven largest, golf-ball-sized xenoliths (18.6–138.8 g with ≤20% host-rock tuff; SLCX 10, 11, 12, 21, 25, 26 and 47) were selected for heavy mineral extraction. To minimize the potential for contamination, pyroxenites were “crushed” using pulse disaggregation with a Lab 1 selfFrag instrument at the Queen's Facility for Isotope Research (QFIR, Queens University, Kingston, ON, Canada). The selfFrag sample vessel, chamber and electrode assembly were fastidiously cleaned before and between samples following a sample-blank bracketing procedure that included: (1) rinsed with tap water, (2) hand washed with mild detergent and a stiff plastic brush, (3) rinsed with tap water, (4) the vessel loaded with a Grenville age marble and processed through the selfFrag instrument (120 pulses at 150 kV), (5) the rinse and washing procedure was repeated before loading each sample, and (6) for each sample, new brushes and plastic containers were used during cleaning to prevent the possibility of cross-contamination. Samples underwent initial, “gentle” disaggregation using 3 L of reverse osmosis water from Collagen Water with 40 pulses at 130 kV to remove fine-grained tuff from the xenoliths. After repeated cleaning, the remaining coarse xenolith material (sieved for >1.4 mm) received an additional 120 pulses at 150 kV to break-up the xenoliths along grain boundaries. The selfFrag instrument does not conventionally “crush” samples, so zircons mantled by other minerals create grains that might not have been liberated during pulse disaggregation. Samples were processed in numerical order (SLCX 10, 11, 12, 21, 25, 26 and 47). Based on records of the samples processed at the QFIR selfFrag facility it is practically impossible that the recovered zircons are “crushing” contaminants. Records show that just before these xenolith samples, multiple Himalayan samples went through the lab during a large three-week project (~10–30 Ma). Other samples pulse disaggregated were from Archean granites, roughly 3.2 to 3.0 Ga, from the Bundelkhand massif in India, and apatite, not zircon, was extracted. Further, no samples with ages similar to those obtained from the xenolith zircons were processed by the QFIR selfFrag for a year prior to the first Oahu xenolith disaggregation to when the last xenolith was processed.

The garnet pyroxenite “xenolith” portion of the seven disaggregated samples was put through magnetic and heavy liquid separation at the Jack Satterly Geochronology Laboratory at the University of Toronto (U of T). After fastidious cleaning of a Franz magnetic separator (water, distilled water and methanol), before and between samples, they went through magnetic separation to obtain nonmagnetic fractions. Steps involved a free-fall removal of magnetic minerals, and then two passes through the separator. The nonmagnetic fraction was put through heavy liquid separation with fastidious cleaning of glassware in an ultrasonic bath before each sample. The seven samples were processed in numerical order (SLCX10

to SLCX47) and seven zircons recovered, four from sample SLCX 12 (3rd sample processed and 2nd smallest at 20.6 g) and three from SLCX47 (last sample processed and largest at 138.8 g).

Only one project went through the Jack Satterly Laboratory (U of T) over the previous 20 years that involved zircon as young as those measured here. This was a study of detrital zircon in the Puna Plateau in the Central Andes⁵³. Mineral processing was done at least a year before this Hawaiian project. Three samples from the Andes project contained a minor component of late Cenozoic zircon. Most of the zircon in all the samples was much older and there was only one zircon found whose age resembles the 14 Ma zircons (rims) in this Hawaiian study (see Fig. 4 in ref. ⁵³).

The zircons cannot be from the xenolith-host material because the volcanoclastic rocks are ≤0.6 Ma old Hawaii¹⁷. Finally, samples were processed in numerical order (SLCX10 to SLCX47) at both Queens and U of T but no zircons were recovered in the first two samples processed (SLCX 10, SLCX 11) nor from the 4th, 5th, and 6th samples, but zircons were recovered from the third (SLCX 12) and 7th (last) xenolith processed (SLCX 47). The recovery of groups of zircons in two samples with extensive cleaning before and between all seven samples and with the rims on all zircons yielding the same ~14 Ma age is consistent with them coming from SLCX 12 and SLCX 47. All of the above makes it extremely improbable that the zircons are from laboratory contamination.

Isotopic analytical methods at the University of Toronto. All LA-ICPMS and TIMS U-Pb ages in Fig. 1 (main text) were plotted in IsoplotR⁵⁴. A ²³⁸U/²³⁵U ratio of 137.88 was used for ²⁰⁷Pb/²⁰⁶Pb model age calculations. U decay constants ($\lambda^{238}\text{U}$ and $\lambda^{235}\text{U}$) come from ref. ⁵⁵ and are $1.55125 \times 10^{-10}/\text{yr}$ and $9.8485 \times 10^{-10}/\text{yr}$. ϵ_{Hf} and Hf model ages were calculated using the equations of ref. ⁵⁶ assuming an age of 13.9 Ma (from U/Pb) and a protolith ¹⁷⁶Lu/¹⁷⁷Hf ratio of 0.013 typical of felsic crust.

Laser ablation, inductively coupled plasma mass spectrometry (LA-ICPMS) analyses of U and Pb were carried out on the seven zircon grains. They were partially ablated using a 213 nm New Wave laser with beam diameter of 10 μm at 5 Hz and about 5 J/cm² power. Mass spectrometry was carried out on a VG Series 2 Plasmaquad equipped with an S-option 75 l/s rotary pump for enhanced sensitivity. Data were collected on ⁸⁸Sr (10 ms), ²⁰⁶Pb (30 ms), ²⁰⁷Pb (70 ms), ²³²Th (10 ms) and ²³⁸U (20 ms). Immediately prior to each analysis, the spot was pre-ablated over a larger area than the beam diameter for about 10 s to clean the surface. Following a 10 s period of baseline accumulation the laser sampling beam was turned on and data collected for 25 s followed by a 50 s washout period. About 150 measurement cycles per sample were produced and ablation pits are about 15 μm deep. Instability was dampened through the use of a 25 ml mixing chamber in-line with the He flow transporting the ablated sample to the plasma.

Data were edited and reduced using custom software written by one of the authors (D. Davis). No corrections were made for common Pb, because this peak is too small to be measured precisely. Common Pb, if present, would have the effect of shifting data to the right away from the concordia curve. ⁸⁸Sr was monitored from zircon in order to detect beam intersection of alteration zones or inclusions. The Th/U ratio of zircon can be a useful petrogenetic indicator and was also measured, although it is only a rough estimate because the ratio is not constant in the standard. The zircon standard was DD91-1, a monzodiorite from the Pontiac province of Quebec previously dated at 2682 ± 1 Ma by isotope dilution thermal ionization mass spectrometry⁵⁷. Sets of 3 sample measurements are bracketed by measurements on standards. Differences between standards are time interpolated to correct sample measurements.

Pb and U isotopes in zircon were also analyzed by chemical abrasion, isotope dilution - thermal ionization mass spectrometry (CA-ID-TIMS) in the Jack Satterly Geochronology Laboratory at the University of Toronto. Zircon grains were placed in a muffle furnace at ~900 °C for ~48 h and etched using 50% Hf in Teflon dissolution vessels at 200 °C for 9 h to remove radiation-damaged and altered zones⁵⁸. Each zircon grain was cleaned in HNO₃ and transferred to a Teflon bomb⁵⁹. Weights were estimated from grain measurements using photomicrographs. A mixed ²⁰⁵Pb-²³³U isotopic tracer solution from the EARTHTIME Project was added to the Teflon dissolution capsules during sample loading (see www.earth-time.org). Zircon was dissolved using ~0.10 ml concentrated HF acid and ~0.02 ml 7 N HNO₃ at 200 °C for 5 days, and redissolved in ~0.15 ml of 3 N HCl. Uranium and lead were isolated from the zircon solutions (reserved for Hf isotopic analysis) using anion exchange columns, deposited onto out gassed rhenium filaments with silica gel⁶⁰, and analyzed with a VG354 mass spectrometer using a Daly detector in pulse counting mode. Corrections to the ²⁰⁶Pb-²³⁸U and ²⁰⁷Pb/²⁰⁶Pb ages for initial ²³⁰Th disequilibrium in the zircon data have been made assuming a Th/U ratio in the magma of 4.2. All common Pb was assigned to a procedural Pb blank. Dead time of the measuring system for Pb was 16 ns and 14 ns for U. The mass discrimination correction for the Daly detector is constant at 0.05% per atomic mass unit. Amplifier gains and Daly characteristics were monitored using the SRM 982 Pb standard. Thermal mass discrimination corrections are 0.10% per atomic mass unit for Pb and U fractionation was measured internally and corrected for each block. Age calculations were done using an in-house program by D.W. Davis. All age errors quoted in the text and table, and error ellipses on the concordia diagram (Fig. 3) are given at the 95% confidence interval.

Zr-Hf-REE bearing column wash solutions from U-Pb CA-ID-TIMS chemistry were collected, dried and redissolved in 1 N HCl + 0.1 N HF. The REE were removed using cation exchange columns. Hafnium isotopes were measured using a Neptune-Plus at the University of Toronto. The instrument was equipped with Jet cones for enhanced sensitivity. ^{172}Yb and ^{175}Lu were monitored to correct any residual interferences from these elements on ^{176}Hf . No correction was made for ^{176}Lu decay, which is negligible because of the young age of the samples.

Clinopyroxene and garnet analyses at the University of Windsor. Clinopyroxene and garnet analyses used an Agilent 7900 fast-scanning quadrupole inductively coupled plasma mass spectrometer coupled with a Photon Machines Excite 193 ultra-short pulse argon fluoride excimer laser (LA QICP MS) at U. Windsor's, Great Lakes Institute for Environmental Research (GLIER). The QICP QMS operated at a RF = 1250 W. Carrier gas (Ar), cell gas (He) and sampling arm gas (He) flow rates were 0.8, 0.36 and 0.84 L/min (respectively). Laser fluence, rep rate, spot size, raster rate, background count time and laser-on count time were 3.46 J/cm², 20 Hz, 25 μm , 5 $\mu\text{m/s}$, 60 s and ~60 s, respectively. Raw data were processed in IOLITE and laser-on signal selection avoided anomalies in all elements across the 60 s analysis time. NIST SRM 610 served as an external standard⁶¹ and internal standard Si concentrations came from electron microprobe analyses. Silicon and Ti, and the other major elements (not reported here) were determined using wavelength dispersive spectroscopy with a Cameca SX5 Field Emission-Electron MicroProbe (EMP) at the UBC Okanagan, Fipke Laboratory for Trace Element Research (FILTER). Operating parameters were 15 keV accelerating voltage, 20 nA beam current, and a 10 μm beam size. Calibration used Smithsonian reference minerals. Precision and accuracy were estimated from replicate analyses of Smithsonian USNM 111356 Hb for Ti and USGS BCR-2G for the LA QICP MS elements. Precision is ~2% for Ti, better than 5% for most elements, 5–10% for Sm, Gd, Dy, Er, and Yb, and ~10% for Lu. Accuracy for the majority of elements is better than $\pm 2.5\%$.

Data availability

All referenced data sets are available in the manuscript (Table 1), in Supplementary Data 1 and online at the data archive <https://doi.org/10.5683/SP2/W0FIPU>.

Code availability

No computer code was written in preparing the paper.

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Author contributions

J.D.G. conceived of the experiment and wrote manuscript drafts from coauthor comments. S.L.K. identified zircon, did the U–Pb TIMS analyses and wrote related text. D.W.D. did the LA ICP MS and Hf analyses, provided results/text and drafted the Papua New Guinea discussion. K.L. did the thermal modeling and finalized the U–Pb dating diagram. Z.Z. wrote draft petrographic descriptions. D.L.-M. did dis-aggregations and contributed associated text. J.D.V. and B.B. helped with Hf analyses.

Competing interests

The authors declare no competing interests.

Additional information

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