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Zircon xenocrysts from Easter Island (Rapa Nui) reveal hotspot activity since the middle Jurassic

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Key Points:

- Presence of ~0-165 Myr old zircons with coherent O and $\epsilon\text{Hf}(t)$ mantle isotopic signatures in the Easter plume.

36 • A large population of ~165 Ma zircons could signal a massive melting phase associated
37 with formation of a Large Igneous Province (LIP)

38 • This LIP probably subducted below the Antarctic Peninsula around 100-105 Ma likely
39 related to the Palmer Land deformation event.

40

41 Abstract

42 The 0–2.5 Ma volcanism in Easter Island (Rapa Nui) emerges just east of the East Pacific Rise
43 on young (Pliocene, 3–4.8 Ma) ocean floor. Here, we report the finding of mantle-derived zircon
44 grains retrieved from Easter Island beach sands and red soils that are much older than the Easter
45 Island volcanism and its underlying lithosphere. A large population of 0–165 Myr old zircons
46 have coherent oxygen ($\delta_{18}\text{O}_{(\text{zircon})}$ 3.8– 5.9‰) and hafnium mantle isotopic signatures ($\epsilon\text{Hf}_{(t)}$
47 +3.5–+12.5). These results are consistent with the crystallization of zircon from plume-related
48 melts. In addition, a chemically diverse population with ages as old as Precambrian was also
49 found. We thus suggest that the Easter hotspot started at least ~165 Ma ago. A large population
50 of ~160-164 Ma zircons could signal an intense initial massive melting phase associated with the
51 formation of a Large Igneous Province (LIP) upon the first arrival of the plume. We use plate
52 reconstructions to show that such a LIP would have formed on the Phoenix Plate. It would have
53 subducted below the Antarctic Peninsula around 100-105 Ma, offering a solution for the
54 enigmatic Palmer Land deformation event, previously proposed to result from a collision with an
55 unknown indenter. Our findings show that asthenospheric mantle-derived xenocryst zircon
56 cargo, as recently reported from Galápagos, may not be an exception. The here-reported “ghost”
57 of a prolonged hotspot activity suggests that the Easter hotspot and the sub-lithospheric mantle in
58 which it is entrained remained mantle-stationary for at least 165 Ma.

59 1 Introduction

60 Plate tectonics theory straightforwardly explains how oceanic lithosphere is subductable and
61 only becomes a few hundred million years old, whereas continental lithosphere is not, and may
62 become billions of years old. Geochronological databases (U-Pb in zircons) derived from
63 continental crust contain age spectra that cover almost all of Earth’s history (e.g. Condie et al.,
64 2009; Domeier et al., 2018; Rodriguez-Corcho et al., 2022; Wu et al., 2023), whereas oceanic
65 crusts tend to only contain zircons in a narrow age spectrum that coincides with the magmatic
66 age of the crust (e.g. Lissenberg et al., 2009; Rioux et al., 2016). In that context, it is puzzling
67 that an increasing body of literature reports zircon populations collected at mid-oceanic ridges
68 and from remote oceanic hotspot volcanoes that are far older than the crust they reside in.

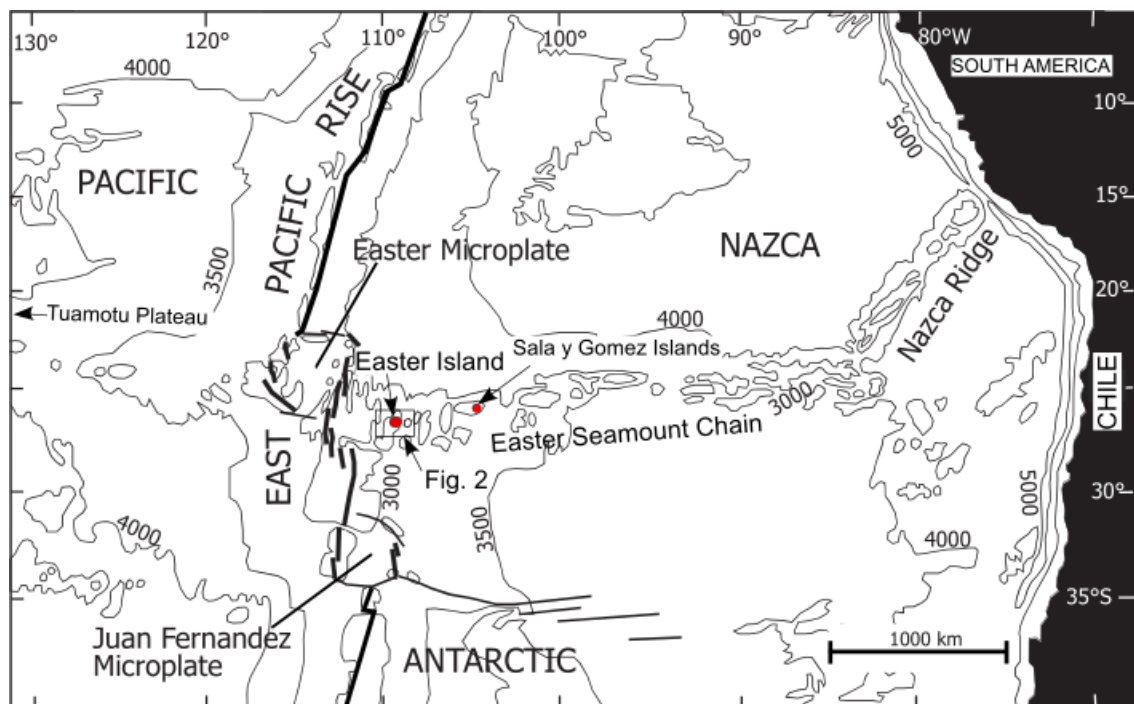
69 Pilot et al. (1998) was the first to report Paleozoic and older zircons from the Central Atlantic
70 mid-ocean ridge. Since this first finding, diverse populations of Paleozoic and older zircons have
71 also again been collected from the Central Atlantic Ridge and reported from the North Atlantic
72 and South Atlantic ridges (Bea et al., 2020; Bjerga & Pedersen, 2021; Bortnikov et al., 2008;
73 Skublov et al., 2022), and recently from intra-oceanic, remote hotspot islands of Hawai’i
74 (Greenough et al., 2021) and Galápagos (Rojas-Agramonte et al., 2022). These ancient zircons
75 cannot have formed from crystallizing magma at the young mid-ocean ridges, or in the oceanic
76 crust below the hotspot volcanoes, and were consequently interpreted as xenocrysts.

77 In addition, ancient zircon populations were found with narrow age ranges or coherent
78 chemistry. For instance, ~180 Ma zircons were collected from the Southwest Indian Ridge
79 (Cheng et al., 2016) and a ~0–164 Ma population with coherent chemistry was recently reported

80 from Galápagos (Rojas-Agramonte et al., 2022). That zircon can survive asthenospheric mantle
81 conditions in mafic and ultramafic systems and retain its U-Th-Pb and Lu-Hf isotopic signature
82 was shown in laboratory experiments (Bea et al., 2018; Cambeses et al., 2023). Furthermore,
83 xenocrystic zircons are routinely found in ophiolitic mantle sections (e.g. Aitchison et al., 2022;
84 González-Jiménez et al., 2017; Lian et al., 2020; Moghadam et al., 2022; Portner et al., 2011;
85 Proenza et al., 2018; Xiong et al., 2022), and subducted zircons resurface in arc volcanoes (Gao
86 et al., 2022; Rojas-Agramonte et al., 2017; Rojas-Agramonte et al., 2016). The serendipitous
87 findings of zircons have thus been explained as ‘ghosts’ (Gianni et al., 2023) of either subducted
88 sedimentary zircons entrained in fossil mantle wedges or detached sub-continental mantle
89 lithosphere that float in the asthenospheric mantle (e.g. Bjerga et al., 2022; Gianni et al., 2023;
90 van Hinsbergen et al., 2020; Pilot et al., 1998) or as magmatic zircons that formed during
91 crystallization of mantle plume-related melts (Rojas-Agramonte et al., 2022). The presence of
92 such features in the sub-lithospheric mantle is consistent with equally fortuitous geochemical
93 findings from Atlantic, Pacific, and Indian mid-ocean ridges (Le Roux et al., 2002; Liu et al.,
94 2022; Richter et al., 2020; Urann et al., 2020; Yang et al., 2021).

95 The preservation of such ghosts could offer novel possibilities to reconstruct the geological
96 history of the mantle. For instance, the finding of a range of zircons from the Jurassic to the
97 present with coherent plume-like chemistries on the Galápagos led Rojas-Agramonte et al.
98 (2022) to postulate that the Galápagos plume may date back far longer than the oldest magmatic
99 products that are typically ascribed to the plume (the Cretaceous Caribbean Large Igneous
100 Province). This suggests xenocrystic zircons on young hotspot islands may offer insight into a
101 far longer geological history of the sub-lithospheric mantle than is preserved on the present
102 oceanic crust.

103 In this paper, we report on the zircon cargo that we collected during a reconnaissance study of
104 beach sands and soils from the remote Easter hotspot island, 3800 km west of the South
105 American coast in the Pacific Ocean (Fig. 1). The Plio-Pleistocene Easter volcanic Island is built
106 on Pliocene ocean floor (~3–4.8 Ma; (Bonatti et al., 1977; Hagen et al., 1990; Naar & Hey, 1989;
107 Vezzoli & Acocella, 2009), and the oldest zircon that could have formed during volcanism or the
108 underlying sea floor may thus be ~4 Ma old. The oldest seamounts that formed at this hotspot
109 that are preserved on the modern ocean floor are in the Tuamotu Seamount chain and are 48 Ma
110 old (Bello-González et al., 2018). The hotspot likely started earlier, but older seamounts have
111 disappeared through subduction below South America or Antarctica, and the onset of plume
112 activity is unknown. We will use the age distributions and composition of the zircons obtained
113 from the Easter beaches and soils to evaluate whether xenocrysts are present and discuss the
114 possibilities that xenocrystic zircon cargo may offer a novel source of information on mantle and
115 plume-related geological history.



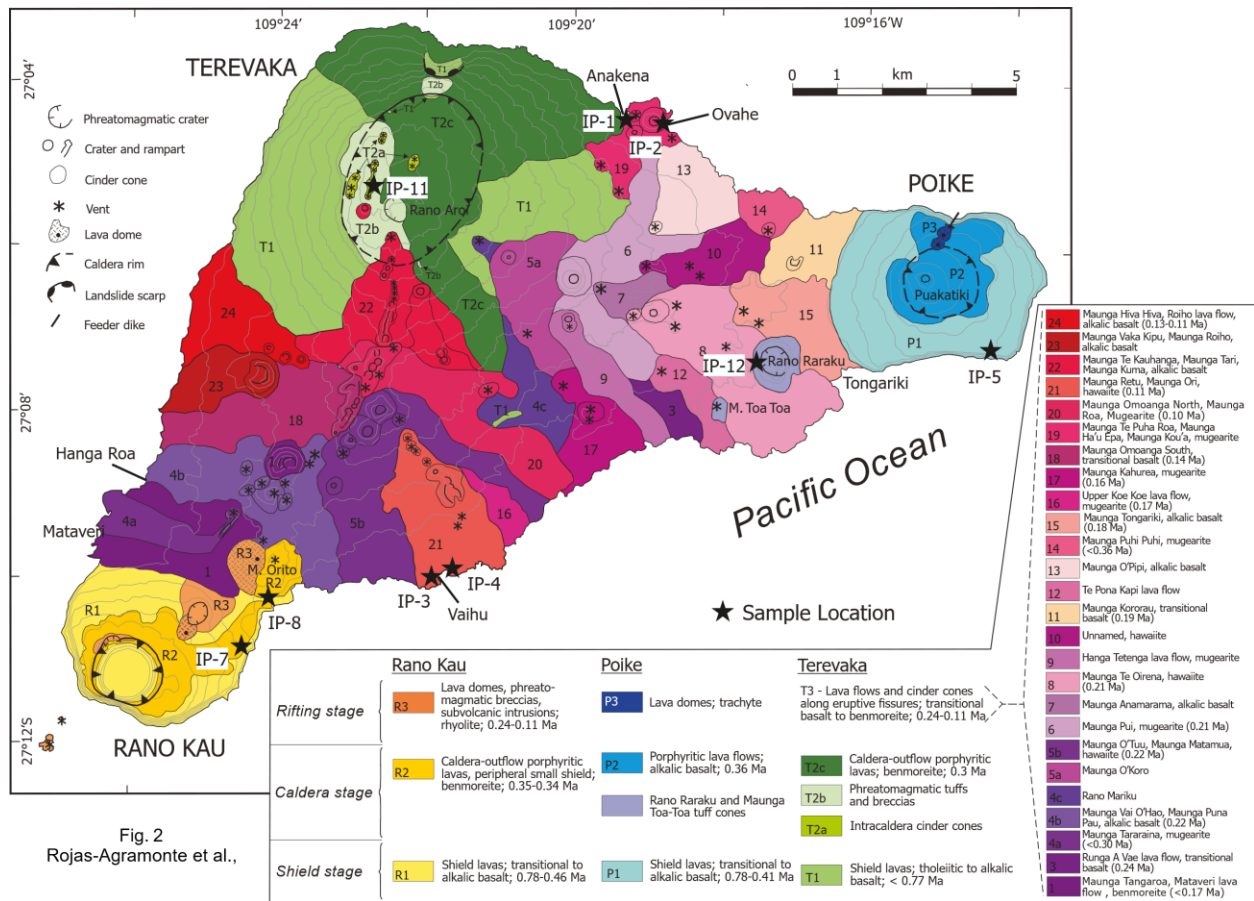
116
 117 *Figure 1. Tectonic realm of the study area showing the location of Easter Island (Rapa Nui) at the westernmost end of the Easter*
 118 *Seamount Chain on the Nazca Plate, east of the East Pacific Rise.*

119 2. Geological background

120 Easter Island (Rapa Nui, Polynesia) lies on the westernmost edge of the Nazca Plate, close to the
 121 East Pacific Rise that forms the spreading ridge with the Pacific Plate (Fig. 1). The Easter
 122 hotspot is located above the edge of the Pacific Large Low Shear Velocity Province (LLSVP),
 123 and is thought to result from a deep mantle plume that originates at ~2900 km depth near the
 124 core-mantle boundary (Burke et al., 2008; Courtillot et al., 2003; Harpp et al., 2014). The
 125 youngest seamount chain is located on the Nazca Plate and forms the 2900 km long, E-W
 126 trending Easter Seamount Chain with the active Easter-Salas hotspot at its western tip (Bello-
 127 González et al., 2018; Ray et al., 2012; Rodrigo et al., 2014; Vezzoli & Acocella, 2009).
 128 Eastwards, the Easter Seamount Chain connects to the Nazca Ridge that represents a ~30–20 Ma
 129 older portion of the chain (Fig. 1) (Bonatti et al., 1977; Rappaport et al., 1997; Ray et al., 2012;
 130 Simons et al., 2002; Stoffers et al., 1994). Between ~48 and 30 Ma, the East Pacific Rise was
 131 located just east of the hotspot, and the Tuamotu Seamount chain formed on the Pacific Plate
 132 (Bello-González et al., 2018; Hampel, 2002; Rosenbaum et al., 2005). Prior to 48 Ma, the
 133 hotspot was presumably located east of the East Pacific Rise (Bello-González et al., 2018; Espurt
 134 et al., 2007) (Fig. 1).

135 The first accounts on the geology and petrography of Easter Island were published by Chubb
 136 (1933), Bandy (1937), and Baker (1967), whilst the first comprehensive geological maps were
 137 provided by Gonzalez-Ferran & Baker (1974), Gonzales-Ferran et al. (2004), and Gioncada et al.
 138 (2010). The island represents the small (160 km²) emerged portion of three main, coalescing

139 polygenetic shield volcanoes (Poike, Rano, Kau, and Terevaka; Fig. 2), built during the last ~2
 140 Ma on top of a ~3–4.8 Ma old oceanic crust (Fig. 2; Bonatti et al., 1977; Hagen et al., 1990; Naar
 141 & Hey, 1989; Vezzoli & Acocella, 2009). The emerged products of these three volcanoes are
 142 mainly effusive, strongly affected by marine erosion, and are crosscut by multiple (~100),
 143 aligned monogenetic vents (Baker, 1967; Gioncada et al., 2010; Vezzoli & Acocella, 2009).
 144 Volcanic products are predominantly basalts, hawaiites, mugearites, and benmoreites, but there
 145 are also trachytes and peralkaline rhyolites (Fig. S1; Baker et al., 1974; Bonatti et al., 1977;
 146 Boven et al., 1997; Clark & Dymond, 1977; DePaepe & Vergauwen, 1997; Haase et al., 1997).
 147 The lavas have a high zirconium content ranging from 137 ppm in the mafic lavas to as high as
 148 1355 ppm in the intermediate and felsic products, which is somewhat not surprising that we have
 149 found so many zircons in the detritus of the island (Table S1). Vezzoli & Acocella (2009)
 150 identified a general evolution pattern for the island volcanism, starting with a shield stage
 151 followed by a non-explosive caldera stage truncating each of the three main edifices, and
 152 culminating with the rifting stage of the island, dominated by monogenetic volcanism.



153
 154 *Figure 2. Geological map of Easter Island (Rapa Nui), integrating previous data of Vezzoli and Acocella (2009) and Gioncada et al. (2010), and showing sampling sites (IP) of this study. Notice that information regarding the three main polygenetic volcanoes (Poike, Rano Kau, and Terevaka) is separated from data related to small-volume monogenetic vents.*

157 Poike Volcano (370 m asl) is located at the eastern end of the island (Fig. 2). Most of the edifice
 158 consists of a thick succession of lava flows exposed along the sea cliffs (Gioncada et al., 2010).

159 K/Ar geochronology dating of the volcano yielded ages ranging from 2.54 ± 0.28 Ma to
160 0.26 ± 0.21 Ma (Baker et al., 1974; Clark, 1975; Kaneoka & Katsui, 1985). Rano Kau (300 m asl)
161 is located at the southwestern end of the island and hosts a distinctive non-explosive central
162 caldera currently filled with a lake (Fig. 2). K/Ar Ages from this volcano range from 0.94 ± 0.19
163 Ma to 0.34 ± 0.03 Ma (Clark, 1975; Miki et al., 1998) (Fig. 2), although field relationships
164 suggested that volcanism here also started around 2.5 Ma (Gioncada et al., 2010). The Terevaka
165 shield volcano (511 m asl) is located at the northern end of the island (Fig. 2). Most of the edifice
166 consists of lava flows whereas the rifted summit is cross-cut by several NNE-SSW aligned
167 craters (Rano a Roi the largest) on top of topographic ridges formed by coalescing and
168 superposed scoria cones (Baker, 1967; Baker et al., 1974; Vezzoli & Acocella, 2009). (Gonzales-
169 Ferran et al., 2004) considered Terevaka volcanism to span between ~ 1.9 and 0.3 Ma from field
170 relationships, the latter relying on a K/Ar age of ~ 0.30 Ma reported by (Baker et al., 1974)
171 without analytical data. Younger K/Ar ages of 0.14 ± 0.06 Ma and 0.12 ± 0.05 Ma were reported
172 by (Clark, 1975) and belong to the late rifting stage of the central edifice (Vezzoli & Acocella,
173 2009).

174 The monogenetic vents (Fig. 2) formed during the rifting stage include (i) two NE-SW aligned
175 phreatomagmatic basaltic tuff cones (P. E. Baker, 1967; Bonatti et al., 1977; Heyerdahl &
176 Ferdon, 1961), (ii) seven evolved vents, comprising three trachyte-rhyolitic lava-domes, four
177 obsidian lava-domes and an obsidian-breccia vent strongly aligned through NE-SW parallel
178 structures and (iii) multiple NE-SW, NNE-SSW and NNW-SSE trending basaltic scoria cones
179 cross-cutting the island (see electronic appendix in Gioncada et al., 2010 and in Vezzoli &
180 Acocella, 2009).

181 The extrusion of evolved magmas is structurally controlled along the NE-SW faults crosscutting
182 the southern polygenetic centers (Haase et al., 1997). The undated trachyte-rhyolite end-member
183 cutting the northern slope of Poike include three lava domes (Maunga Vai a heva, Maunga Tea-
184 tea, and Maunga Porehe; Baker, 1967; Baker et al., 1974; Bonatti et al., 1977). On the
185 northeastern slopes of Rano Kau, there is an effusive and an explosive vent. The former is the
186 well-known Maunga Orito lava-dome, which was K/Ar dated between 0.34 ± 0.06 Ma and 0.24
187 ± 0.03 Ma (Clark, 1975; Heyerdahl & Ferdon, 1961; Kaneoka & Katsui, 1985; Miki et al., 1998;
188 Vezzoli & Acocella, 2009). The explosive vent corresponds to the Te Manavai crater, which is
189 the source of a distinctive spherulitic, obsidian-rich tuff breccia (Baker, 1967). Vezzoli &
190 Acocella (2009) reported K/Ar ages of 0.11 ± 0.04 Ma and 0.18 ± 0.003 Ma, which likely
191 correspond to fragments derived from this explosive source. In addition, the NE-SW Motu-nui,
192 Motu-iti, and Motu-kaokao islets, which emerged to the southwest of the island, also expose
193 rhyolites and obsidian (Vezzoli & Acocella, 2009).

194 Dry basaltic, hawaiite, and mugearite scoria cones are the most common monogenetic vents
195 across the island, following NNE-SSW, NNW-SSE, and NE-SW structural trends (Gioncada et
196 al., 2010, in Gonzales-Ferran et al., 2004). The documented K/Ar ages for the monogenetic

197 volcanism range from 0.24 ± 0.05 Ma to 0.09 ± 0.02 Ma (James Gregory Clark, 1975) without
198 showing a specific trend in vent migration (see appendix in Vezzoli & Acocella, 2009).

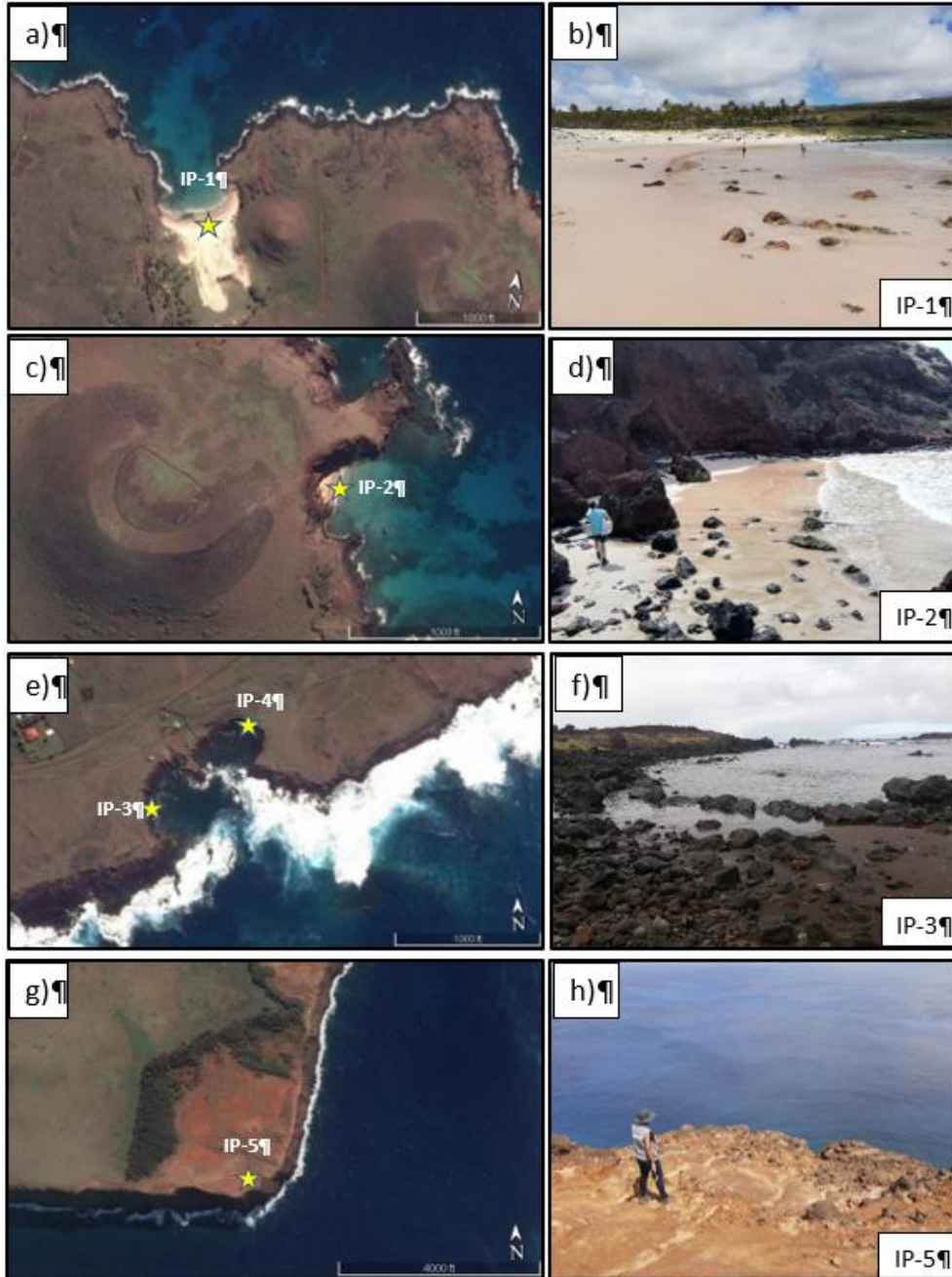
199 **3 Sampling, methods, and analytical procedures**

200 During the reconnaissance field trip that lies at the basis of this paper, rock sampling was not
 201 allowed due to geoh heritage restrictions. A total of twelve (12) detrital samples were collected,
 202 four from beach sands and eight from soil in the island's interior (Fig. 2). Of these, nine samples
 203 contain zircon grains (Table 1; Figs. 2 and 3). The sampling procedure included the selection of
 204 representative localities where sufficient soil and sand were available and that covered different
 205 potential source areas from the volcanic vents.

206 *Tabelle 1 Sampling sites 'with coordinates and general description.*

Sample	Coordinates		Locality	Lithology	Comment
	Latitud	Longitud			
IP-1	S 27° 04' 23"	W 109° 19' 20"	Anakena beach	Beach sand	Distinct by its predominant bioclastic componentry compared to the rest of the studied localities
IP-2	S 27° 04' 26"	W 109° 18' 57"	Obahe beach	Beach sand	On the eastern flank of the scoria cones belonging to the polygon 19 in the Vezzoli and Acocella (2009) map.
IP-3	S 27° 08' 52"	W 109° 24' 59"	Vaihu beach	Beach sand	The samples are the product of erosion from regoliths on top of lava flows belonging to the NNW-SSE vents of the polygon 21 mapped by Vezzoli and Acocella (2009). Precense of pahoehoe lava-flows, locally showing pillow structures.
IP-4	S 27° 09' 56"	W 109° 21' 44"	Kona Ha'ari - Otongariki	Beach sand	Idem to IP-3
IP-5	S 27° 07' 12"	W 109° 14' 35"	Poike Volcanoe	Red soil	The red lateritic soil developed on top of the P1 aphyric to porphyritic lava flow succession forming the southeastern flanks of Poike shield volcano and exposed along the high sea cliffs
IP-7	S 27° 10' 45"	W 109° 24' 22"	Vinapu	Red soil	Red soil profile/regolith developed on top of a succession of bedded lava-flows exposed on the southeastern flanks of Rano Kau shield volcano. The sample overlays the unit mapped as R1 at Maunga Orito.
IP-8	S 27° 10' 12"	W 109° 24' 03"	On the road	Red soil	Red soil profile/regolith developed on top of a succession of bedded lava-flows exposed on the southeastern flanks of Rano Kau shield volcano. The sample overlays the unit mapped as R2, at Maunga Orito
IP-11	S 27° 05' 29"	W 109° 22' 52"	Terevaka Volcanoe	Red soil	Sample collected from the regolith developed on top of the youngest, NNE-SSW aligned scoria cones belonging to the unit mapped by Vezzoli and Acocella (2009) as T2a.
IP-12	S 27° 07' 24"	W 109° 17' 29"	Ranu-Raraku	Soil	Sample collected from the regolith and locally washed epiclastic silts exposed on the western flank of the Rano Raraku tuff cone

207
 208 All samples were panned with seawater directly on the island and only the heavy fraction was
 209 taken. The magnetic fraction of each sample was separated with a neodymium hand magnet and
 210 further panned with ethanol in the laboratory. Zircons for isotopic analysis were then handpicked
 211 during optical inspection under a binocular microscope and mounted in epoxy resin. The mount
 212 was ground down and polished to expose the interiors of the grains. Grains were photographed in
 213 reflected and transmitted light and under cathodoluminescence (CL; after carbon coating) to
 214 enable easy and best location for analytical spots during analysis. CL imaging was performed in
 215 the EMP (Electron Microprobe) facilities at Mainz University using a JEOL JXA-8200.
 216 Operating conditions were 20.0 kV and 30nA.

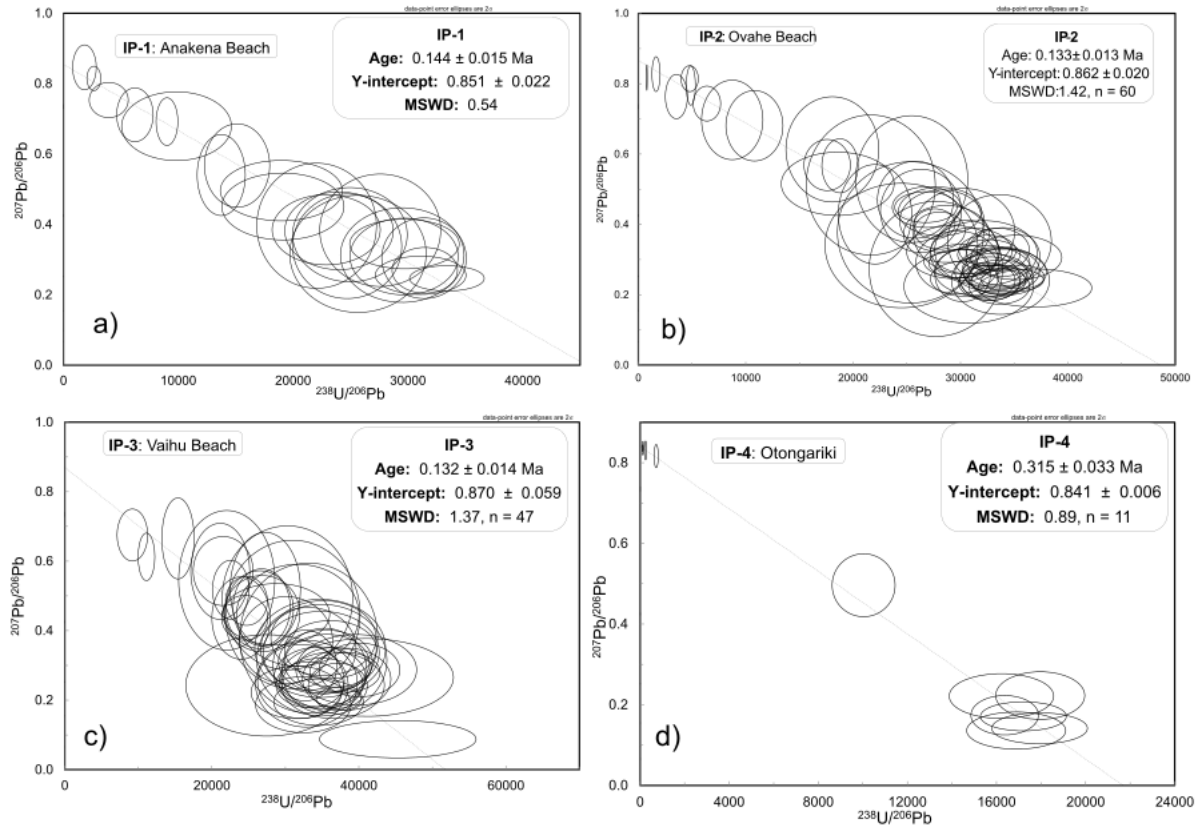


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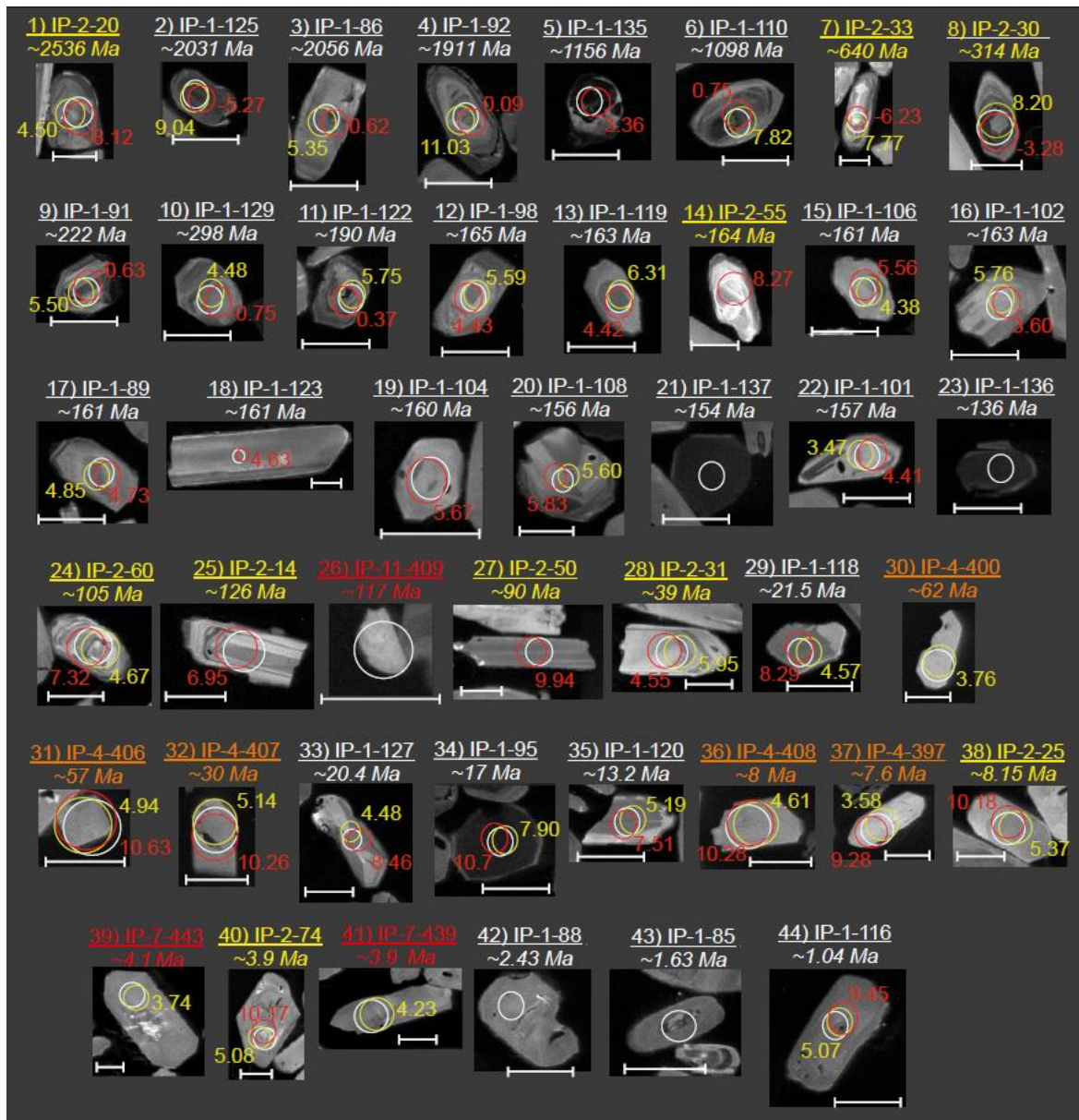
218 *Figure 3 Google Earth images and photos of sample locations IP-1, IP-2, IP-3, and IP-5 (see the geological map on Fig. 2 for*
 219 *sample locations). a) Google Earth image of Anakena Beach, b) Photo of Anakena Beach, c) Google Earth image of Ovahe*
 220 *Beach, d) Photo of Ovahe Beach, e) Google Earth image of Vaihu (left) and Palmeras de Otongariki (right), f) Photo of Vaihu*
 221 *Beach, g) Google Earth image of cliff southeast of Poike volcano, h) photo from red soil on the cliff (150 m) southeast of Poike*
 222 *volcano*

223 Zircon was analyzed in situ for U-Pb and Hafnium (Hf) at the laboratory facilities of the
 224 Frankfurt Isotope & Element Research Center (FIERCE, Germany), for Oxygen (O) at the
 225 SHRIMP II of the Beijing SHRIMP Centre, Chinese Academy of Geological Sciences and for
 226 trace elements at the laboratories of the State Key Laboratory of Lithospheric Evolution, Institute

227 of Geology and Geophysics, Chinese Academy of Sciences. The results appear in Tables S2-S4.
 228 Analytical procedures are presented in detail in the -supporting Information. A correction for
 229 initial ^{230}Th and U-series disequilibrium was applied to all zircon grains younger than 2 Ma (For
 230 more details see Rojas-Agramonte et al., 2017).



231
 232 *Figure 4 Concordia diagrams and U-Pb ages retrieved from zircons sampled in Rapa Nui beaches; a) IP-1 Anakena Beach, b)*
 233 *IP-2 Ovahe Beach, c) IP-3 Vaihu Beach, and d) IP-4 Otongariki site.*



234
 235 *Figure 5 Cathodoluminescence images of zircons retrieved from each sample, marked in colors and organized from oldest (to)*
 236 *youngest (bottom). Regions of interests for in situ analyses are marked as white circles for U-Pb, red circles for hafnium, and*
 237 *yellow circles for oxygen isotopic measurements.*

238 4 Results

239 The sand on each beach in Easter Island is unique in composition, color, and grain size, as a
 240 result of biological activity, their source rocks, and coastal processes. The two main beaches on
 241 the island are located in the northern part (Ovahe and Anakena). White coral sands dominate the
 242 210 m wide, sand-rich embayed beach in Anakena (Figs. 3a, b; sample IP-1). Dark/red volcanic-
 243 derived detritus dominates in the 70m wide Northern pocket beach Ovahe, forming soft sands in
 244 both (Figs. 3c, d; sample IP-2). The sediment-starved beaches of Vaihu in the south are
 245 dominated by dark volcanic-derived blocks and detritus (Figs. 3e, f; samples IP-3 and IP-4).

246 The soils of Easter are all of volcanic origin, derived either from pyroclasts or weathered
 247 volcanic rocks. The soils are red, yellow, or brown in color and in places appear slightly
 248 undulating with steep cliffs on some of the edges facing the sea, especially on the margins of the
 249 Rano Kao and Poike volcanoes (Figs. 3h, h, sample IP-5). Samples IP-7, IP-8, IP-11, and IP12
 250 also come from red soils sampled across the island (Fig. 2). All samples contain a variable
 251 amount of zircon grains except for samples IP-11 and IP-12, where zircon is sparse (Figs. S9 and
 252 S10). Zircons older than the age of the volcanism of the island are concentrated in the north in
 253 samples IP-1 and IP-2 (Fig. 2). In sample IP-1 of 49 dated zircons 24 are old and in sample IP-2
 254 of 76 dated zircons 9 are old (Table S2). In addition, sample IP-4 (contains 5 older zircons out of
 255 16) and IP-5 with only one older zircon out of 75 dated. See detail description below.

256 **4.1.1 In-situ U-Pb in zircon**

257 Analyses were guided by cathodoluminescence (CL) images (Figs. S2 to S10 of the Supporting
 258 Information). The zircon crystals in the studied samples vary in size (from 100 up to 550 μm)
 259 and shape. Although most crystals are euhedral (equant to elongated), some are broken or
 260 rounded. In some cases, the zircons show oscillatory zoning. There is no clear correlation
 261 between zircon shape, size, and age group, although some of the oldest zircons appear with
 262 rounded terminations and clear core and rim features. The rims are too small to be dated by LA-
 263 ICP-MS (e.g. Fig. S2; spot number 92).

264 A total of forty-nine (49) grains were recovered from sample **IP-1** at the enclosed Anakena
 265 Beach (Figs. 3a, b; Table S2). These are mostly Pleistocene in age ($n=24$), between 2.43 ± 0.46
 266 Ma and 0.19 ± 0.02 Ma, yielding a concordant age at 0.144 ± 0.015 Ma (Fig. 4a). One more
 267 grain yielded a Pliocene age (3.5 ± 1.6 Ma) and the rest are significantly older. Four Miocene
 268 (21.5 ± 0.5 to 13.2 ± 0.3 Ma), one Lower Cretaceous (136 ± 7 Ma), eleven Jurassic (190 ± 5 to
 269 154 ± 10 Ma), one Upper Triassic (222 ± 5 Ma), one early Permian (298 ± 6 Ma), three
 270 Mesoproterozoic (1242 ± 20 , 1156 ± 22 , and 1098 ± 21 Ma), and three Paleoproterozoic ($2059 \pm$
 271 32 , 2031 ± 38 and 1911 ± 38 Ma) grains were dated (Figs. 5 and S11).

272 A total of seventy-six ($n = 76$) grains were recovered from sample **IP-2**, at the enclosed Ovahe
 273 Beach (Figs. 3c, d). Most of them ($n = 66$) also provided Pleistocene ages between 1.82 ± 0.43
 274 Ma and 0.17 ± 0.02 Ma, yielding a concordant age at 0.133 ± 0.013 Ma (Fig. 4b). One more
 275 grain is Pliocene (3.9 ± 0.7 Ma) in age, and the rest are significantly older. One Miocene ($8.1 \pm$
 276 0.4 Ma), one Eocene (39.1 ± 0.8 Ma), three Cretaceous (from 90.2 ± 2.0 to 126 ± 3 Ma), one
 277 Middle Jurassic (164 ± 3 Ma), one Middle Carboniferous (314 ± 6 Ma), one Neoproterozoic (640
 278 ± 12 Ma), and one Early Paleoproterozoic (2536 ± 42 Ma) grains were dated (Figs. 5 and S11).

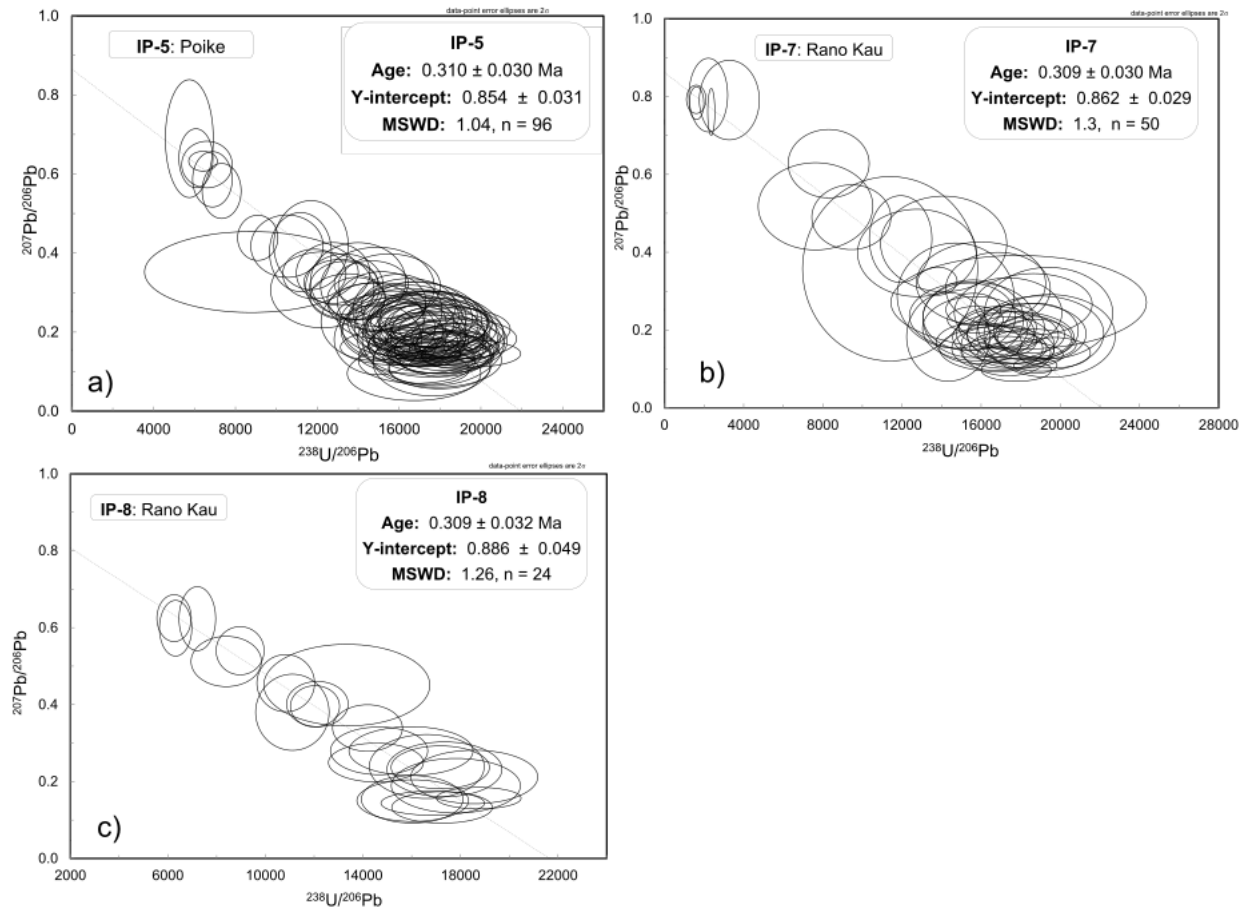
279 We recovered sixty-five (65) grains from sample **IP-3**, at Vaihu beach (Figs. 2 and 3e, f). All of
 280 them consistently provided Pleistocene ages between 0.38 ± 0.05 Ma and 0.11 ± 0.05 Ma,
 281 yielding a concordant age at 0.132 ± 0.014 Ma (Fig. 4c). The neighboring sample **IP-4** provided
 282 fourteen (14) grains. Ten (10) Pleistocene grains dated between 0.58 ± 0.12 Ma and 0.18 ± 0.01
 283 Ma, yielding a concordant age of 0.315 ± 0.033 Ma (Fig. 4d). The other four grains yielded two

284 Miocene ages (8.4 ± 0.6 and 7.6 ± 0.4 Ma) and one Oligocene (29.6 ± 4.8 Ma) and one
285 Paleocene (57.0 ± 9.1 Ma) age (Fig. 5).

286 Sample **IP-5**, from a red soil from Poike volcano (Figs. 3g, h) contains a large amount of zircon
287 grains ($n = 92$). Most of the zircons provided Pleistocene ages between 1.70 ± 0.74 and $0.31 \pm$
288 0.04 Ma, yielding a concordant age of 0.310 ± 0.030 Ma (Fig. 6a) and one Eocene grain ($49.4 \pm$
289 8.7 Ma; Fig. 5).

290 On Rano Kau flanks, most zircons recovered from lateritic soil in sample **IP-7** ($n = 53$), yielded a
291 concordant Pleistocene age of 0.309 ± 0.030 Ma (ranging from 1.60 ± 0.34 to 0.418 ± 0.05) (Fig.
292 6b). Two other grains are Pliocene in age (4.10 ± 0.74 and 3.90 ± 1.05 Ma). In addition, a total of
293 27 grains were recovered from sample **IP-8**, yielding a concordant Pleistocene age of $0.309 \pm$
294 0.032 Ma (Fig. 6c), ranging between 1.184 ± 0.02 Ma and 1.184 ± 0.02 Ma.

295 Very few zircon grains were recovered from red soil at the summit of Terevaka (sample **IP-11**).
296 Only two grains were dated, one being Early Pleistocene ($\sim 2.27 \pm 0.3$ Ma) and the other one
297 Cretaceous (117 ± 50 Ma; Fig. 5). Likewise, most of the heavy minerals recovered from black
298 soil at Rano Raraku (sample **IP-12**) are apatite. The only zircon found was dated at 0.57 ± 0.19
299 Ma.



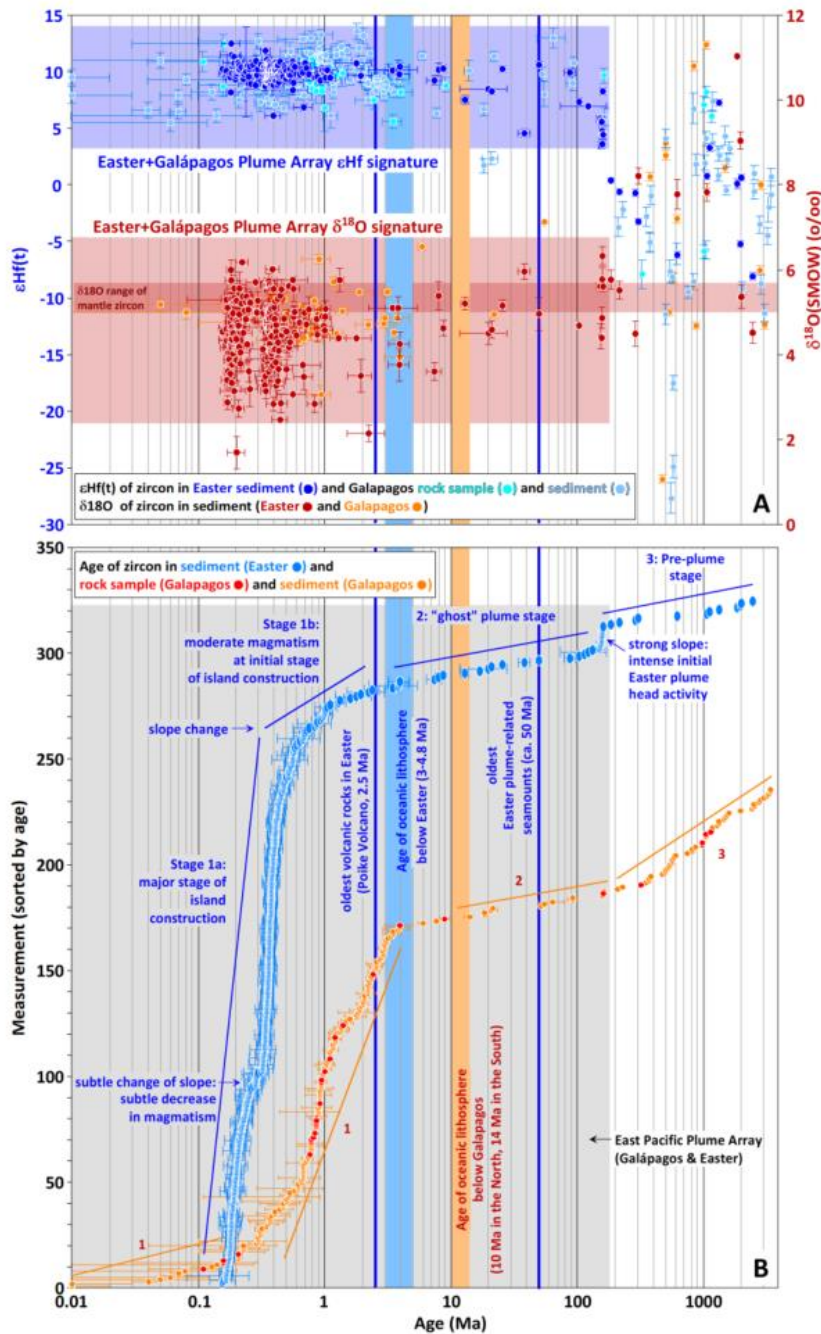
300

301 *Figure 6 Concordia diagrams and U-Pb ages retrieved from zircons sampled on the flanks of the main shield polygenetic*
 302 *volcanoes; a) IP-5 at Poike, b) IP-7 at Rano Kau, and c) IP-8 at the limits between Ran Kau and Mt Orito.*

303 4.1.2 Zircon Lu-Hf and Oxygen isotopes composition

304 The zircon $\epsilon\text{Hf}(t)$ and $\delta_{18}\text{O}_{(\text{zircon})}$ values are plotted versus U-Pb age in Fig. 7A. All zircon grains
 305 with ages ~ 0 to 165 Ma exhibit a range of positive $\epsilon\text{Hf}(t)$ values (+3.5 to +12.5), however, most
 306 of the analysis have an average of $\epsilon\text{Hf}(t) = +8$ to +10 (Figs. 7A and S12; Table S3). A small but
 307 distinct shift towards lower ϵHf compositions is observed as the grains become older compared
 308 to the younger zircon grains with lower values of +3.5 in the Jurassic grains (Fig. S12). Despite
 309 the spread in the ϵHf values these results reflect derivation from a predominantly juvenile
 310 depleted mantle source with the involvement of an isotopic heterogeneity in the source of the
 311 magmas. The oxygen isotope field for ‘mantle’ zircon ($+5.3 \pm 0.3\%$; Valley, 2003) is shown for
 312 reference in Figs 7A and S12. Zircons younger than ~ 165 Ma show a modest range in $\delta_{18}\text{O}$
 313 values between 3.8 and 5.9‰ except for one grain (~ 163 Ma) showing a more elevated value of
 314 6.3‰. Approximately 25% of the zircons investigated plot within or straddle within error the
 315 mantle zircon field. An important feature revealed in these data is that the majority of the zircon
 316 oxygen isotope compositions plot below the mantle zircon reference field and indicate a more
 317 complex history of hydrothermal processes or interaction with high-T seawater in the upper part
 318 of the oceanic lithosphere (Valley, 2003). The isotopic composition of the zircon grains older

319 than 164 Ma (Fig. 8) is characterized by more diverse and variable $\epsilon\text{Hf}(t)$ (-8.1 – +7.26) and
 320 $\delta^{18}\text{O}_{(\text{zircon})}$ (4.4 – 11.3‰) values, consistent with both juvenile and continental crust signatures
 321 (Fig. 7A and S12; Table S4).



322
 323 *Figure 7 Isotopic composition ($\epsilon\text{Hf}(t)$ and $\delta^{18}\text{O}(\text{zircon})$) and age of Easter zircons. The zircon data of Galápagos of Rojas-*
 324 *Agramonte et al. (2022) are also indicated for reference. (A) U-Pb age versus $\epsilon\text{Hf}(t)$ (bluish colors) and $\delta^{18}\text{O}(\text{zircon})$ (reddish*
 325 *colors) of analyzed zircons. The Easter+Galápagos Plume Array is defined by high $\epsilon\text{Hf}(t)$ and low $\delta^{18}\text{O}(\text{zircon})$, which extends*
 326 *from 0 Ma to 165 Ma (note significant scatter for Easter and Galápagos at >165 Ma). (B) Age of analyzed zircons sorted by age*
 327 *of spot. Note sectors with varied slopes denoting variable abundance of zircon ages for the corresponding time sectors.*

328 4.1.3 Zircon trace element composition

329 Chondrite-normalized REE values are plotted by age range (~ 0 –165 Ma and >165 Ma) in Fig.
330 S13 and Table S5. Zircon younger than 165 Ma exhibits extreme enrichment in heavy rare earth
331 elements (HREEs) relative to light rare earth elements (LREEs) and well-developed positive Ce
332 anomalies and negative Eu anomalies (Fig. S13). These compositions are consistent with
333 crystallization from very oxidized (~ 2 to $3 \log_{10}$ orders $fO_2 >$ synthetic nickel-nickel oxygen
334 buffer) magmas in which Eu^{2+} was extremely limited in abundance (Burnham and Berry, 2012;
335 Trail et al., 2012). Uranium and Thorium abundances range from ~ 37 to ~ 1340 ppm and ~ 17 to
336 ~ 1890 ppm, respectively (Table S5). Despite this wide range of concentrations, they form a
337 well-defined array with a Th/U ratio that varies systematically from ~ 0.2 to ~ 2 with increasing U
338 and Th. Hafnium concentrations are mainly between ~ 940 and ~ 15000 ppm. Zircons older than
339 165 Ma are less abundant and show REE patterns that display chondrite-normalized trends with
340 enrichment in HREEs relative to LREEs and positive Ce and a less pronounced negative Eu
341 anomalies compared to the younger grains (Fig. S13). Uranium and Thorium abundances display
342 a narrower range compared to the younger zircon, ranging from ~ 134 to ~ 356 ppm and ~ 13 to
343 ~ 122 ppm, respectively and a lower Th/U ratio ranging from 0.09 to 0.7 (Table S4; S5).

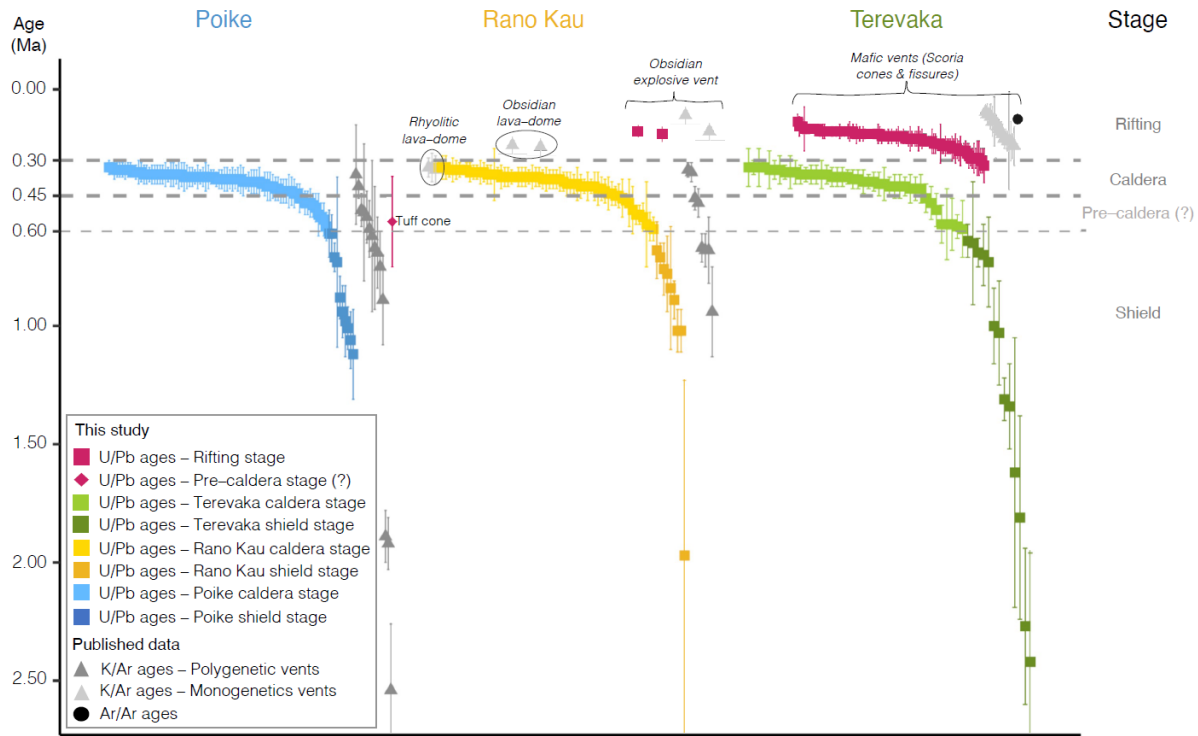
344 5 Discussion

345 5.1. Plio-Pleistocene ages

346 The Pliocene-Pleistocene ages constitute the vast majority (80%) obtained in this study (Figs. 8
347 and S14). They are within the range of K/Ar ages reported in the literature and show that zircon
348 also crystallized from the magma that formed the volcano. Our results suggest that magmatism at
349 Rapa Nui has been nearly continuous between 3.9 ± 0.74 and 0.11 ± 0.05 Ma. Pleistocene
350 zircons retrieved from samples collected on the flanks of the main polygenetic volcanoes (IP-5,
351 IP-7, and IP-8) consistently yielded concordant ages of ~ 0.3 Ma, whereas samples collected
352 closer to the youngest monogenetic and fissure vents and products (IP-1, IP-2, and IP-3) yielded
353 concordant ages at ~ 0.13 – 0.14 Ma.

354 After correlating our data with previous ages, the magmatism that yielded the construction of the
355 main shields spanned from ~ 4.0 to ~ 0.6 Ma. Afterward, we identify a marked increase of ages
356 between ~ 0.6 and ~ 0.3 Ma corresponding to the magmatism related to the caldera-stage
357 volcanism identified by Vezzoli & Acocella (2009). Finally, magmatism driving the
358 monogenetic and fissural volcanism cross-cutting the island likely started at ~ 0.51 Ma and is
359 continuous down to 0.14 ± 0.03 Ma (Fig. S14). All data from monogenetic vents were treated
360 separately because these sources are not necessarily related to the same reservoir-plumbing
361 system as the three main edifices (Fig. 8). The oldest zircon age found at a monogenetic vent was
362 retrieved at the Rano Raraku tuff cone (0.56 ± 0.19 Ma; IP-12), but we do not have an
363 independent dating technique to constrain the onset of phreatomagmatism in the island. U/Pb
364 ages between 0.51 ± 0.07 Ma (in IP-3) and 0.33 ± 0.08 Ma (in IP-1) show a good correlation
365 with published K/Ar data for lavas of the early caldera stages at polygenetic volcanoes because

366 our sampling sites correspond to the erosion products of the lavas and scoria cones mapped by
 367 Vezzoli & Acocella (2009). The concordia ages obtained at IP-7, IP-8, and IP4 suggest that dry
 368 mafic monogenetic volcanism was already occurring sometime around 0.309 ± 0.030 Ma and
 369 0.315 ± 0.033 Ma. These results confirm the onset of the rifting stage of the island (Vezzoli &
 370 Acocella, 2009).



371
 372 *Figure 8 Geochronological database of Rapa Nui, summarizing previously published Ar^{40}/Ar^{39} and K^{40}/Ar^{40} ages, compare to our*
 373 *new U/Pb zircon ages (Table S6). Most of the data correspond to shield, pre-caldera, and caldera stages of each polygenetic*
 374 *central volcanoes. Younger ages corresponding to monogenetic vents formed during the rifting stage less represented and shown*
 375 *in detail Figure S14.*

376 Consistently, evolved (rhyolitic) magmatism also started at ~ 0.3 Ma with the emission of
 377 rhyolitic lava-domes (dated by Gonzalez-Ferran and Baker, 1974; Miki et al., 1998) followed by
 378 ~ 0.2 Ma old obsidian lava domes dated by Kaneoka and Katsui (1985). The sampling locations
 379 and U/Pb ages of IP 7 (0.19 ± 0.03 Ma) and IP-8 (0.18 ± 0.02 Ma) clearly correlate with the 0.18
 380 ± 0.03 Ma K/Ar age reported at the Te Manavai obsidian-rich explosive vent by Gonzalez-Ferran
 381 & Baker, 1974). All younger ages retrieved from locations close to scoria cones and the youngest
 382 mapped lava flows yielded ages between 0.30 ± 0.05 Ma (in IP-3) and 0.14 ± 0.03 Ma (in IP-3),
 383 with ~ 0.13 – 0.14 Ma concordant ages clearly correlating to the published data for the latest
 384 eruptions (Vezzoli & Acocella, 2009). Most zircons within this age group have positive ϵ_{Hf}
 385 values between $+6.1$ – 12.5 (Table S3), indicating a predominantly juvenile-depleted mantle
 386 source.

387 The finding of zircons with crystallization ages older than the ~ 3.0 – 4.8 Ma age of the oceanic
 388 lithospheric basement cannot be explained by crystallization of the lavas on the island, or of the

389 oceanic crust below the volcano. They occur mostly in the North (IP-1 and IP-2 with the highest
390 age variability) and Southeast (IP-4). Zircon in the range ~4–165 Ma bear $\epsilon\text{Hf}_{(t)}$ and $\delta^{18}\text{O}$
391 signatures typical of mantle zircon similar to the pre-4 Ma zircons (Figs. 7A and S12), showing
392 that they come from the erosion of the mantle-derived volcanic products. The finding of zircon
393 grains on hotspot-related islands that are much older than the age of volcanism in the island or
394 the underlying oceanic crust (i.e., >4 Ma in the case of Easter Island) has far-reaching
395 geodynamic implications if they were brought up from the underlying mantle as xenocrysts
396 (Rojas-Agramonte et al., 2022). It is therefore important to consider whether the zircon cargo
397 from the beaches and soils may have been transported from elsewhere. Possibilities are then
398 aeolian, marine, or human transport. Aeolian transport should then transport zircons either blown
399 by strong winds or volcanic eruptions.

400 Oceanic currents could transport zircons from the continent to Easter Island, which is located
401 within the South Pacific Gyre, a large ocean current that is also responsible for the transport of
402 plastics from the coast of Peru and Chile to the island (Gennip et al., 2019). Theoretically, old
403 zircons could arrive on the island transported in or on plastic, or as part of large floating pumice
404 rafts created by eruptions of submarine volcanoes (Jutzeler et al., 2020; Ohno et al., 2022).
405 Finally, old zircons could have been transported to the island as ship ballast, or otherwise
406 through human transport.

407 While contributions from these sources cannot be entirely excluded, it is hard to explain the
408 zircon age distribution and chemistry with these alternative sources. The distance of Easter
409 Island to the South American continent of ~3800 km makes airborne transport of zircon crystals
410 with sizes of several hundred μm (Figs. S2 and S3) very unlikely. Even if the zircons would have
411 been introduced through pumice rafts, the question remains how these old zircons ended up in
412 the pumice. The coherent chemistry of the zircons of the last 165 Ma does not have an obvious
413 source in South America, where they would have much more diverse isotopic signatures
414 (signaling Andean volcanism, e.g., Balgord, 2017; Pankhurst et al., 2016) than those found in the
415 present study. A key test to demonstrate that the zircons are derived from the volcano is the
416 finding of a zircon cargo within the lavas that are comparable to those of the soils and beach
417 sands. Such confirmation that sands and lavas from hotspot islands carry similar zircon cargo
418 was shown elsewhere, on Galápagos (Rojas-Agramonte et al., 2022), and Mauritius (Ashwal et
419 al., 2017; Torsvik et al., 2013). Nonetheless, natural processes such as erosion and detrital
420 sediment accumulation, sorting, and temporary and past deposition are efficient ways to
421 concentrate zircons in oceanic island environments (Rojas-Agramonte et al., 2022; Rojas-
422 Agramonte et al., 2017; Seelos et al., 2021; Sevastjanova et al., 2011). Bearing in mind that we
423 have not been able yet to demonstrate the *in-situ* presence of zircons xenocrysts in the Easter
424 Island lavas, we will in the remainder of this discussion explore the implications of our findings
425 assuming that they represent xenocrysts from the sublithospheric mantle below the volcano.
426 Besides the hot-spot effusive eruptions typical of polygenetic shield volcanoes, monogenetic
427 volcanism is known to be an excellent feeder mechanism able to drag deep accessory and

428 accidental grains from their asthenospheric source to the surface (Brenna et al., 2011; R. Gao et
429 al., 2017; Simon et al., 2008).

430 Early Jurassic (~190 Ma), Triassic (~222 Ma), Carboniferous (~314 Ma), and Precambrian (~640
431 to 2500 Ma) zircon ages retrieved at Rapa Nui are not expected in the oceanic crust. Most of the
432 zircon grains older than 165 Ma have values ranging from 0 to negative ϵ_{Hf} isotopic signatures
433 typical of continental magma sources or of contamination with continental crustal material. Only
434 two grains point toward a juvenile source (Fig. 7A). Such ancient zircons with diverse
435 compositions were also reported from Galápagos (Rojas-Agramonte et al., 2022) and Hawai'i
436 (Greenough et al., 2021) and are best explained as detrital zircons that were brought into the
437 mantle by subduction. Whereas Galápagos is located in the vicinity of a subducted slab in the
438 underlying lower mantle that subducted in the Mesozoic (the Malpelo slab, Van der Meer et al.,
439 2018) no slabs are known below Hawaii and Easter Island. If these zircons resulted from
440 subduction, they were either transported over large distances by mantle flow (Greenough et al.,
441 2021), or they were introduced into the mantle when continents were still nearby. The
442 geochemistry of central Pacific hotspots reveals geochemical traces of subducted continental
443 crust that is widely thought to be present in the plume source at the core-mantle boundary and
444 that subducted in the late Precambrian to early Paleozoic (e.g., Jackson & Macdonald, 2022)
445 This would suggest that these xenocrysts may have been trapped in the upper mantle for
446 hundreds of millions of years (Greenough et al., 2021; van Hinsbergen et al., 2020; Rojas-
447 Agramonte et al., 2022), likely enclosed within mineral grains that provides a way to remain
448 stable under asthenospheric conditions (Bea et al., 2020; Cambeses et al., 2023).

449 The ~165–0 Ma zircon, with clear Hf and O isotopic mantle signatures and well-covered age
450 range from ~0–165 Ma appears to date the production of magmas for an extended period of time
451 in the location of the current hotspot. The most straightforward explanation is that the current
452 volcano is located at the site of a long-lived hotspot that dates back to the Middle Jurassic (~165
453 Ma). The pre-Pliocene zircons then likely represent minerals that formed and were stored in the
454 asthenosphere, and that were picked up at sublithospheric depths by rising hot-spot magmas.
455 Interestingly, this age range of hotspot-related zircons is similar to the one reported from
456 Galápagos (Rojas-Agramonte et al., 2022), suggesting that both hotspots may date back to the
457 middle Jurassic.

458 Figure 7B clearly shows different stages that we interpret as plume development and oceanic
459 island formation of Easter Island compared to the recent findings in the Galapagos Archipelago.
460 The younger stages are the island-formation events, with stage 2 which is the most recent and
461 vigorous, denoted by the steep slope. This stage is characterized by the abundance of young
462 zircons (<1 Ma), related to the majority of exposed rocks that are subject to erosion. Stage 2a is
463 older and less vigorous (<3 Ma; not so many zircons for a larger time interval), as denoted by the
464 less steep slope of the curve. Stage 1, which we dub the "ghost" plume stage, contains the
465 xenocrysts and ends with the formation of oceanic lithospheres on which the modern islands are
466 built. It is characterized by a fewer occurrence of zircons during the long period of time of plume

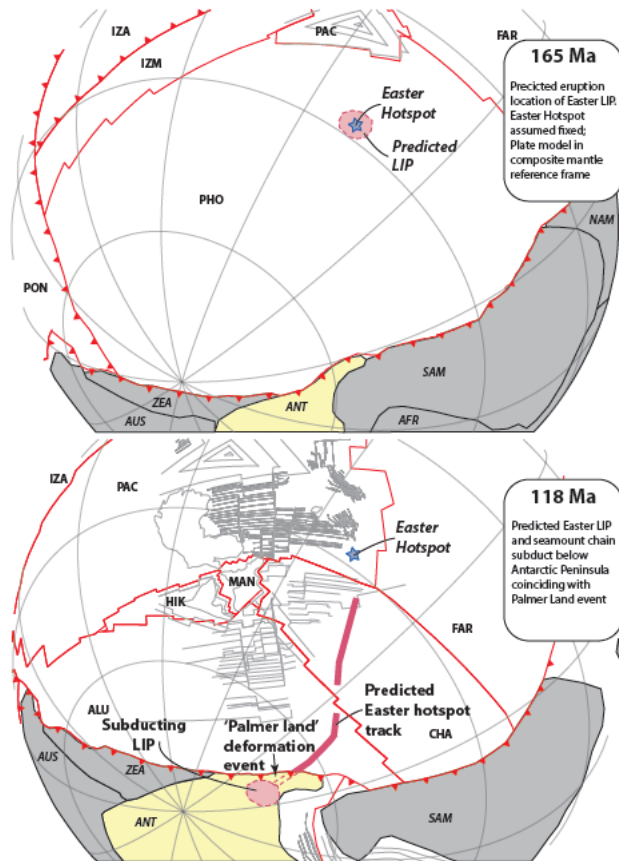
467 activity (Fig. 7B: gentle slope). The gentle-slope curve extends to ages younger than the
468 corresponding lithosphere, but these zircons do not correspond to stage 2 island-forming lavas
469 zircons, but to older asthenospheric or lithospheric zircons recovered from lavas erupted later
470 during stage 2. This “ghost” stage appeared to have abruptly started in the Jurassic with a strong
471 slope. corresponding to a relative abundance of zircons ($n = 8$) for a relatively short time interval
472 (156–165 Ma) at the beginning of the EPPA. We postulate that this was related to the first arrival
473 of the plume head below the lithosphere, which could then have generated a Large Igneous
474 Province (LIP).

475 The lithosphere that was present above the present-day locations of the Easter and Galápagos
476 hotspots has long been lost to subduction, but we perform an experiment to evaluate where the
477 hypothetical, ~165 would have subducted. We will use this to test whether a geological record of
478 this LIP, or of the effects of its subduction, may still be present in circum-Pacific orogens.

479 The arrival of LIPs in trenches is known to have major tectonic effects. The arrival of the
480 Ontong-Java plateau at the Melanesian trench caused strong upper plate deformation that even
481 led to subduction polarity reversals (e.g. Knesel et al., 2008). However, accretion of crustal units
482 from LIPs is rare, probably because there are few horizontal decollement horizons in the dyke-
483 pierced LIPs, and they eventually tend to entirely subduct (Van Hinsbergen & Schouten, 2021)
484 or only leave minor accreted ocean plate stratigraphy thrust slices (van de Lagemaat et al., 2023).
485 Hence, we may evaluate whether, around the time interval in the location of predicted
486 subduction, there is a record of upper plate shortening.

487 For our reconstruction (Fig. 9), we use a recently updated plate model of the southern Pacific
488 region back to the early Cretaceous (van de Lagemaat et al., 2023), and a simple plate model
489 based on mirroring the Early Cretaceous and Late Jurassic magnetic anomalies preserved in the
490 Pacific plate prior to that (Boschman et al., 2021). The relative motions between the Pacific plate
491 and its oceanic neighbors and the Pangea-derived continents of the Indo-Atlantic plate system is
492 directly constrained through a connection to Antarctica for the last 83 Ma (e.g. Torsvik & Cocks,
493 2019). However, prior to this connection, the Pacific Ocean plates and the surrounding
494 continents were everywhere separated by trenches or transforms, and relative motions between
495 these plate systems relies on independent mantle reference frames. For the Pacific plate, we use
496 the latest fixed hotspot frame of Torsvik et al. (2019), which extends back in time to 150 Ma, and
497 for the Indo-Atlantic plate system we use the slab-fitted reference frame of that is the only one
498 that goes back to the Jurassic. (Boschman et al., 2019) showed that these two reference frames
499 perform well in predicting paleomagnetic and age grid constraints for the transition of the
500 Caribbean plate from the Pacific to the Indo-Atlantic plate systems around 100 Ma. Before 150
501 Ma, we use the approximate position of the Pacific and neighboring oceanic plates relative to the
502 Indo-Atlantic plates that was estimated by Boschman et al., 2021) that was chosen such that
503 subduction is maintained at all trenches surrounding the Panthalassa ocean. The uncertainty in
504 the position of the Pacific relative to the Indo-Atlantic plates is difficult to quantify, but is likely

505 on the order of 500–1000 km. Hence, the absolute ages that we estimate below should not be
 506 taken too literally, but the uncertainty has little effect on the region and the general timing of
 507 subduction of the LIPs.



508
 509 *Figure 9 A) Paleogeographic reconstructions of the southern Pacific realm in a composite mantle reference frame (see text),*
 510 *indicating the predicted location of the Easter LIP at 165 Ma, assuming that the Easter hotspot has remained mantle-stationary.*
 511 *B) The plate model predicts that the 165 Ma-old Easter LIP would have subducted around 118 Ma below the Antarctic*
 512 *Peninsula, followed by the subduction of an associated seamount chain until ~108 Ma. This coincides with the enigmatic Palmer*
 513 *Land deformation event on the Antarctic Peninsula.*

514 For our reconstruction, we assume that the Galápagos and Easter hotspots have remained
 515 stationary in the mantle since 165 Ma. We then draw a circular LIP around the location of the
 516 hotspot at 165 and attach it to the plate that was at that time located above the hotspot. For the
 517 Galápagos hotspot, this was the Farallon plate, likely just north of the Farallon-Phoenix ridge,
 518 and for the Easter hotspot, this was the Phoenix Plate. The Galapagos LIP would likely have
 519 been located in the lithosphere that was consumed by Early Cretaceous subduction below the
 520 intra-oceanic subduction zone whose remains are found in the northern Caribbean region, from
 521 Guatemala over Cuba to Hispaniola. The Cuban ophiolites have been intensely deformed after
 522 the Early Cretaceous (e.g. Iturralde-Vinent et al., 2016), and while Early Cretaceous upper plate
 523 deformation may well have occurred, there is no possibility to confirm or exclude this scenario.

524 Our reconstruction shows that the hypothetical Easter LIP would have subducted below the
525 Antarctic Peninsula, starting around 120 Ma, and a hypothetical seamount chain would have
526 subducted below the Peninsula until ~110–105 Ma (Fig. 9). After that, the seamount chain would
527 have subducted below South America. Interestingly, this coincides with the enigmatic ‘Palmer
528 Land’ deformation event (Vaughan et al., 2002). This deformation comprises NW-SE shortening
529 that started at or before 116 Ma, and that ended around 106 Ma, by which time the deformed
530 rocks were unconformably covered by undeformed arc sequences (Vaughan et al., 2002, 2012).
531 Vaughan et al. (2002, 2012) inferred that this deformation was likely the result of the arrival of a
532 buoyant indenter at the trench, but no accretionary geological record has so far been recognized.
533 The subduction of a 165 Ma Easter LIP and following would provide a straightforward
534 explanation for this poorly understood shortening event.

535 It is early days to use zircon xenocrystic cargo as basis for geodynamics, and there are many
536 sampling targets in the oceans that may provide a wealth of new information. Nonetheless, we
537 consider our test of whether zircon cargo from remote oceanic islands can offer a new
538 geodynamic perspective to be positive. Our findings provide a second example of what we
539 hypothesize is the ghost of a prolonged plume history, and the coincidence of the predicted LIP
540 subduction and the Palmer Land deformation event is striking. If our hypothesis is valid, the
541 geodynamic implications are far reaching. It would suggest that not only the Galápagos and
542 Easter hotspots are nearly mantle-stationary, but also that the asthenosphere in which the
543 underlying plume is contained did not appreciably move for the last 165 Ma. Because if the
544 mantle would convected and moved laterally with a plume fed from below continuously piercing
545 ‘fresh’ asthenosphere, the older melt relics and zircon cargo would have convected away.
546 Previously, geochemical traces of subduction found in young volcanic rocks that were unrelated
547 to subduction were linked to detached slabs in the underlying mantle, suggesting that former
548 ‘ghost’ mantle wedges are still located above the slabs that made them, 10’s to 100’s of Ma after
549 slab breakoff (Gianni et al., 2023; van Hinsbergen et al., 2020; Richter et al., 2020). The finding
550 of two 165 Ma long plume records below Galápagos (Rojas-Agramonte et al., 2022) and Easter
551 Island (this paper) provides further arguments towards this surprising finding. Whereas we
552 consider our findings preliminary, and more data and experiments will likely be needed to
553 unlock the full potential of mantle-derived zircon xenocrysts, we conclude that our study
554 provides ample reason to build a much larger database of intra-oceanic volcanic island zircon
555 cargo.

556 **6 Conclusions**

557 Recent serendipitous findings of zircon xenocrysts in intraplate volcanic islands such as Hawai’i
558 and Galápagos led us to test whether such a cargo is common and may be used as a novel source
559 of information for geodynamics. Here we report zircons that we collected from beach sands and
560 soils of Easter Island (Rapa Nui), a remote hotspot island 3800 km west of the South American
561 mainland. Approximately 80% of these zircons returned U/Pb ages that coincide with the
562 previously constrained ages of volcanism and the age of the underlying oceanic crust. However,

563 20% is considerably older and falls in two populations. One population of zircons with a spike of
564 165 Ma and covering the 165–4 Ma interval has coherent oxygen ($\delta_{18}\text{O}_{(\text{zircon})}$) and hafnium
565 mantle isotopic signatures ($\varepsilon\text{Hf}_{(t)}$), consistent with crystallization of zircon from plume-related
566 melts. An older population dates back to the Archean and has diverse chemistries that point to a
567 continental source. We postulate that the older zircons may relate to phases of subduction that
568 occurred when continents still occupied this region in Paleozoic and older times, and suggests
569 that zircon xenocrysts may survive in the asthenosphere for 100s of Ma, echoing previous
570 conclusions. The ~165–4 Ma population would be straightforwardly explained by crystallization
571 from melts generated by the Easter Hotspot, which would then have initiated ~165 Ma ago, long
572 before the oldest, ~48 Ma relics identified in the geological record. We postulate that the
573 population of ~165 Ma zircons could signal an intense initial massive melting phase associated
574 with the impingement of the plume at the base of the lithosphere forming a Large Igneous
575 Province (LIP). Using plate reconstructions, we show that this hypothetical LIP would have
576 formed on the Phoenix Plate and subducted beneath the Antarctic Peninsula between ~120 and
577 105 Ma. This coincides with the enigmatic Palmer Land deformation event that started before
578 116 Ma, and ended around 106 Ma, previously proposed to result from the collision of an
579 unknown indenter. Our findings show that asthenospheric mantle-derived xenocryst zircon
580 cargo, such as that recently reported from Galapagos and Hawai'i, may not be an exception. The
581 "ghost" of prolonged hotspot activity described here suggests that the Easter hotspot as well as
582 the sublithospheric mantle in which it is located remained stationary for some 165 Ma, which, if
583 true, has major implication for our understanding of mantle convection. Even though we consider
584 our results preliminary, we conclude that zircon geochronology combined with elemental (REE)
585 and isotopic studies of young oceanic islands may offer a novel source of information that could
586 upset our thinking of global-scale geodynamics.

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597

598 **Open Research**

599 All the underlying data needed to understand, evaluate, and build upon the reported research are supply in the
600 supporting information.

601

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