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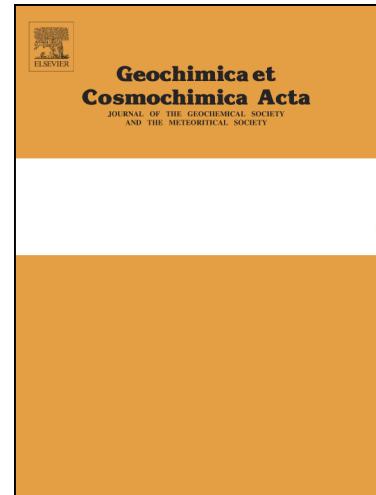
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Rapid Cenozoic ingrowth of isotopic signatures simulating “HIMU” in ancient lithospheric mantle: Distinguishing source from process

Alex J. McCoy-West¹*, Vickie C. Bennett¹ and Yuri Amelin¹

¹Research School of Earth Sciences, Australian National University, Canberra, 2601, Australia

* Present address: Department of Earth Sciences, Durham University, Durham, DH1 3LE, UK

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Corresponding author: Alex McCoy-West (alex.mccoywest@gmail.com)

1 **ABSTRACT**

2 Chemical and isotopic heterogeneities in the lithospheric mantle are increasingly being
3 recognised on all scales of examination, although the mechanisms responsible for generating
4 this variability are still poorly understood. To investigate the relative behavior of different
5 isotopic systems in off-cratonic mantle, and specifically the origin of the regional southwest
6 Pacific “HIMU” (high time integrated $^{238}\text{U}/^{204}\text{Pb}$) Pb isotopic signature, we present the first
7 U-Th-Pb, Rb-Sr, Sm-Nd and Re-Os isotopic dataset for spinel peridotite xenoliths sampling
8 the subcontinental lithospheric mantle (SCLM) beneath Zealandia. Strongly metasomatised
9 xenoliths converge to a restricted range of Sr and Nd isotopic compositions ($^{87}\text{Sr}/^{86}\text{Sr}$
10 =0.7028-0.7033; $\epsilon_{\text{Nd}} \approx +3\text{--}+6$) reflecting pervasive overprinting of their original melt depletion
11 signatures by carbonatite-rich melts. In contrast, rare, weakly metasomatised samples possess
12 radiogenic Nd isotopic compositions ($\epsilon_{\text{Nd}} > +15$) and unradiogenic Sr isotopic compositions
13 ($^{87}\text{Sr}/^{86}\text{Sr} < 0.7022$). This is consistent with melt extraction at ca. 2.0 Ga and in accord with
14 widespread Paleoproterozoic Re-Os model ages from both weakly metasomatised and the
15 more numerous, strongly metasomatised xenoliths. The coupling of chalcophile (Os), and
16 lithophile (Sr and Nd) melt depletion ages from peridotite xenoliths on a regional scale under
17 Zealandia argues for preservation of a significant mantle keel ($\bullet 2$ million km 3) associated with
18 a large-scale Paleoproterozoic melting event. Lead isotopic compositions are highly variable
19 with $^{206}\text{Pb}/^{204}\text{Pb} = 17.3\text{--}21.3$ ($n = 34$) and two further samples with more extreme
20 compositions of 22.4 and 25.4, but are not correlated with other isotopic data or U/Pb and
21 Th/Pb ratios in either strongly or weakly metasomatised xenoliths; this signature is thus a
22 recent addition to the lithospheric mantle. Lead model ages suggest that this metasomatism
23 occurred in the last 200 m.y., with errorchrons from individual localities providing ages
24 younger than 116 Ma. When considered in the regional tectonic context the Pb isotopic
25 signatures are best explained through interaction of the lithospheric mantle with a weak

26 upwelling mantle plume that contained carbonatitic domains at ca. 110-115 Ma. Projection of
27 the measured high U/Pb and Th/Pb signatures into the future predicts extreme Pb isotopic
28 values distinct from any recognised terrestrial reservoir. We suggest that this type of young,
29 carbonatite-related radiogenic Pb signature with extreme $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$, which is
30 widely observed in the southwest Pacific, may reflect a secular change in mantle chemistry
31 consistent with the increased prevalence of carbonatite sources during the Phanerozoic. This
32 signature is referred to as “CarboHIMU”, to differentiate it from the originally defined HIMU
33 representing an ancient lower mantle component present in some ocean island basalts.

34

35 Keywords: Sr-Nd-Pb isotopes; HIMU; carbonatite; New Zealand; mantle xenoliths.

36 **1. INTRODUCTION**

37 Debate continues on the origin of isotopic signatures preserved in the lithospheric mantle as
38 sampled by xenoliths, particularly as to the relationship between siderophile and chalcophile
39 (Pb, Os) isotopic signatures, which may be carried largely in trace phases such as sulfides and
40 metal alloys and lithophile (Sr, Nd) isotopic signatures that are hosted in major phases, which
41 may be overprinted by crustally derived fluids. Understanding the mechanisms responsible for
42 generating the observed isotopic heterogeneities and decoupling of isotopic systems, has
43 widespread implications for mantle evolution as well as significance for how we interpret
44 ages of lithospheric mantle events as determined from different isotopic systems. Related to
45 this are key questions of if, and to what degree, a genetic link exists between signatures in the
46 lower mantle and sub-continental lithospheric mantle (SCLM) and under what conditions
47 these signatures are transferred.

48 In the southwest Pacific, Zealandia, the broader New Zealand micro-continent, is a
49 largely submerged continental ribbon ($>3.5 \times 10^6 \text{ km}^2$) formed during terrane accretion at the
50 eastern margin of Gondwana during the Phanerozoic from ca. 520-100 Ma. Present-day New
51 Zealand comprises a complex collage of geological terranes (Fig. 1), of which the Cambrian
52 to Early Cretaceous basement is divided into two major provinces (Mortimer, 2004): the older
53 early Paleozoic Western Province composed of Gondwanan continental foreland, and the
54 younger (late Paleozoic to mid-Cretaceous) Eastern Province that comprises predominantly
55 deformed meta-greywackes that were accreted during a series of collisional events. These two
56 provinces are separated by the Median Batholith, a long-lived arc-root plutonic complex that
57 records the history of subduction related volcanism from the Carboniferous to Late
58 Cretaceous at the Pacific margin of Gondwana (Mortimer et al., 1999). Intraplate basalts
59 erupted sporadically throughout Zealandia over the last 100 m.y. are notable for their large
60 range ($^{206}\text{Pb}/^{204}\text{Pb} = 18.2\text{-}20.9$) and some highly radiogenic Pb isotopic ratios ($^{206}\text{Pb}/^{204}\text{Pb}$

61 >20.5; McCoy-West et al., 2010; Panter et al., 2006; Timm et al., 2010) and generally
62 unradiogenic $^{87}\text{Sr}/^{86}\text{Sr} < 0.704$. Their SCLM source has been variously referred to as
63 containing HIMU, HIMU-like or enriched mantle components.

64 Here we present new coupled U-Th-Pb, Rb-Sr, and Sm-Nd isotopic measurements
65 combined with published Re-Os isotopic evidence obtained from the same samples (McCoy-
66 West et al., 2013) for a regional suite of mantle xenoliths from throughout Zealandia
67 including the Chatham Islands (Fig. 1). The goals of this study are, firstly, to investigate the
68 relative behaviours of the chalcophile and lithophile element isotopic systems in recording
69 lithosphere formation and modification events in this off-cratonic region, and, secondly, to
70 use integrated geologic and isotopic observations to determine the origin and age of the
71 widespread radiogenic Pb “HIMU” isotopic signature observed in the SCLM and derivative
72 basalts throughout the southwest Pacific.

73 As originally defined, the HIMU (high time integrated $^{238}\text{U}/^{204}\text{Pb}$ or high- μ ; Zindler
74 and Hart, 1986) mantle end-member represents the component with the most radiogenic Pb
75 isotopic signatures ($^{206}\text{Pb}/^{204}\text{Pb} > 20.5$) observed in OIB (ocean island basalts) and with
76 relatively unradiogenic $^{87}\text{Sr}/^{86}\text{Sr} (< 0.703)$. OIB with HIMU signatures are rare; the classic
77 localities of St Helena in the Atlantic Ocean and some of the Cook-Austral islands (e.g.
78 Mangaia, Tubuai and Rurutu) in the Pacific Ocean, although being characterised by high
79 $^{206}\text{Pb}/^{204}\text{Pb}$, also have low $^{208}\text{Pb}/^{206}\text{Pb}$ ratios, similar to those of MORB (mid-ocean ridge
80 basalts), requiring low time-integrated Th/U ratios and high time-integrated U/Pb and Th/Pb
81 ratios, unlike those of any other mantle components (Stracke et al., 2005). Various
82 mechanisms have been proposed for generating HIMU signatures in the mantle including
83 early core formation, delamination of the SCLM and metasomatism by CO₂ rich fluids (e.g.
84 McKenzie and O’Nions, 1983; Meijer et al., 1990; Nakamura and Tatsumoto, 1988; Vidal and
85 Dosso, 1978), although more recently there has been a consensus that recycled ancient

86 oceanic lithosphere is involved in their formation (e.g. Chauvel et al., 1992; Hofmann, 2003;
87 Hofmann and White, 1982; Stracke et al., 2003; Stracke et al., 2005; Weaver, 1991;
88 Woodhead, 1996). Quantitative modelling has shown that modification of the oceanic
89 lithosphere during subduction leads to substantial Pb loss compared to both U and Th and a
90 preferential leaching of U relative to Th (Kelley et al., 2005; Stracke et al., 2003). Thus
91 following subduction modification it is possible to generate a source with the appropriate U-
92 Th-Pb ratios that following extended storage (ca. 0.5-3 Ga) in the lower mantle will develop
93 HIMU characteristics (Chauvel et al., 1992; Hofmann and White, 1982; Stracke et al., 2003).
94 While some authors still use HIMU in the originally defined sense (e.g. Stracke et al., 2005),
95 it is increasingly common to use the term “HIMU” or HIMU-like to describe all basalts with
96 $^{206}\text{Pb}/^{204}\text{Pb} > 19.5$ without full consideration of the genetic implications.

97

98 2. METHODS

99 Details of the acid leaching, chemical separation and mass spectrometry procedures are
100 provided in the Electronic Annex and summarised here. Rubidium-Sr, Sm-Nd and U-Th-Pb
101 concentrations and isotopic compositions were measured at the Australian National
102 University. Whole rock powders were prepared using an agate ring mill, with clinopyroxene
103 separates obtained from larger xenoliths using standard mineral separation techniques. Prior
104 to digestion, all samples were sequentially leached in dilute HNO_3 and hot 6M HCl using a
105 protocol comparable to that recommended by Wittig et al. (2009). Following acid leaching,
106 mixed ^{85}Rb - ^{84}Sr , ^{149}Sm - ^{150}Nd spikes and a synthetic ^{233}U - ^{236}U - ^{229}Th - ^{202}Pb - ^{205}Pb double spike
107 (Amelin et al., 2010) were added to the samples. This was followed by HF- HNO_3 digestion in
108 sealed Teflon beakers on a hotplate at 130 °C. After the samples were completely dissolved,
109 U-Th and Pb were separated using dilute HBr-anion exchange chromatography following the
110 methods of Amelin (2008). Rubidium, Sr and the rare earth elements (REE) were then

111 sequentially separated from the remaining solution using cation exchange chromatography.

112 Strontium was further purified using Sr-spec resin, and Sm and Nd were separated from the
113 REE fraction using Ln-Spec resin.

114 Strontium, Nd, Sm and Pb isotopic compositions were determined by static collection
115 on Faraday cups using a Triton Plus thermal ionisation mass spectrometer (TIMS). Rubidium
116 isotopic compositions were measured in static collection mode on a MAT 261 TIMS.

117 Reproducibility of reference standards throughout the analytical campaign (2σ population)
118 was as follows: 10-100 ng aliquots of SRM987 $^{87}\text{Sr}/^{86}\text{Sr} = 0.710242 \pm 17$ ($n = 12$); 50-100 ng
119 aliquots of an in-house Nd standard GSC AMES $^{143}\text{Nd}/^{144}\text{Nd} = 0.511977 \pm 10$ ($n = 15$) and
120 $^{149}\text{Sm}/^{152}\text{Sm} = 0.516853 \pm 19$ ($n = 11$); 1 ng of SRM981 $^{206}\text{Pb}/^{204}\text{Pb} = 16.940 \pm 18$,
121 $^{207}\text{Pb}/^{204}\text{Pb} = 15.497 \pm 13$ and $^{208}\text{Pb}/^{204}\text{Pb} = 36.711 \pm 55$ ($n = 13$). Uranium and Th
122 concentrations were measured using an Aridus II desolvating nebuliser coupled to a Neptune
123 Plus multi-collector ICP-MS. Analytical blanks in this study were in all cases negligible with
124 no corrections applied. Replicate digestions of rock standard BCR-2 are within error of
125 published Sr-Nd-Pb isotopic values; isotopic compositions of peridotite standards DTS-1 and
126 PCC-1 are in good agreement with the limited published values (Table EA1).

127

128 3. SAMPLES AND RESULTS

129 Mantle xenoliths were collected from 13 localities from throughout Zealandia, including the
130 North, South and Chatham Islands (Fig. 1). Petrographic descriptions, mineral chemistry,
131 whole rock major and trace element concentrations including the platinum group elements and
132 Re-Os isotopic data for all of these samples are available in McCoy-West et al. (2013; 2015).

133 The xenoliths are predominantly harzburgites or clinopyroxene-poor lherzolites and are
134 exclusively from the spinel-facies with equilibration temperatures varying between 830 and
135 1020 °C (McCoy-West et al., 2015). They represent a variably depleted portion of off-

136 cratonic SCLM with Al_2O_3 varying from 0.1-3.4 wt % that has undergone 3-28% melting (F_{Yb}
137 $_{\text{in WR}}$; McCoy-West et al., 2015). Unradiogenic Os isotopic compositions (McCoy-West et al.,
138 2013) for a geographically restricted subset of xenoliths (the Waitaha domain; Fig. 1) that
139 have Re depletion model ages and Re-Os age relationships in accord with widespread melt
140 depletion at ca. 1.9 Ga have established the existence of a large region (>2 million km^3) of
141 Paleoproterozoic SCLM beneath New Zealand (McCoy-West et al., 2015; McCoy-West et al.,
142 2013). In contrast, samples from other regions of Zealandia are characterised by having
143 heterogeneous Os isotopic compositions and Re-depletion ages and lacking correlations on
144 Re/Os versus $^{187}\text{Os}/^{188}\text{Os}$ diagrams (Liu et al., 2015; McCoy-West et al., 2013). Strongly
145 metasomatised xenoliths dominate the xenolith collection, with trace element signatures of
146 clinopyroxene having been interpreted to reflect widespread interaction with a carbonatite
147 component in the SCLM under Zealandia (McCoy-West et al., 2015; Scott et al., 2014a;
148 2014b).

149 **3.1. The effect of acid leaching on Sr-Nd-Pb isotopes**

150 Numerous studies of mantle xenoliths have demonstrated that owing to the ubiquitous
151 presence of exogenic Pb combined with the low concentrations (i.e. sub 10 ppb) in
152 unmetasomatised mantle materials, leaching is required to achieve meaningful U-Th-Pb ratios
153 and Pb isotopic compositions (Wittig et al., 2009). A wide range of leaching protocols have
154 previously been implemented (Hamelin and Allègre, 1988; McDonough and Chauvel, 1991;
155 Pearson et al., 1993; Wittig et al., 2009). While leaching procedures using an HF step can
156 result in severe fractionation of U-Th-Pb ratios, more moderate leaching using only HCl and
157 HNO_3 as employed here, has not been shown to cause disturbance of parent-daughter ratios
158 (see review in Wittig et al., 2009). After leaching, residues measured by TIMS typically
159 contain comparable concentrations to the average of 8-14 unleached crystals measured by
160 laser ablation (McCoy-West et al., 2015; Fig. 2). As seen in previous studies, leaching of

161 clinopyroxene removes a significant fraction (up to 99% of the Pb), with most of this being
 162 exogenous anthropogenic Pb. In this study, the proportion of the U, Th and Pb removed
 163 during leaching varies significantly (ca. 5-99 %; although it is a strong function of the
 164 concentration of the element in the sample; Table EA3; Fig. EA2). Parent-daughter ratios for
 165 the majority of samples plot on a 1:1 line (U/Pb; Sm/Nd; Fig. 2), showing that leaching prior
 166 to TIMS analysis has not fractionated their parent-daughter ratios. The Pb isotopic
 167 compositions of the measured leachates and residues also form a mixing line with Broken Hill
 168 Pb (Cooper et al., 1969), the pervasive anthropogenic Pb component seen throughout
 169 Australasia, consistent with the progressive removal of exogenous Pb during leaching (Fig.
 170 EA3). Further consideration of the effects of acid leaching on parent-daughter ratios and
 171 isotopic compositions are documented in the Electronic Annex.

172 **3.2. Sr-Nd isotope data**

173 The host basalts containing these xenoliths range in age from 86-1.8 Ma, with initial isotopic
 174 compositions for each xenolith calculated using the relevant eruption age. Initial $^{87}\text{Sr}/^{86}\text{Sr}$
 175 compositions from all 34 xenoliths vary from 0.7019-0.7042 (excluding sample P43153b:
 176 $^{87}\text{Sr}/^{86}\text{Sr} = 0.7093$; Table 1) with 26 samples having a restricted $^{87}\text{Sr}/^{86}\text{Sr}$ range (0.7026-
 177 0.7032); no correlation is observed with $^{87}\text{Rb}/^{86}\text{Sr}$ (Fig. 3a). Replicate analyses of
 178 clinopyroxene from three fertile, weakly metasomatised samples preserve the most
 179 unradiogenic Sr signature (OU45852; DPP-1 & 5; $^{87}\text{Sr}/^{86}\text{Sr} = 0.7019-0.7021$). Initial
 180 $^{143}\text{Nd}/^{144}\text{Nd}$ compositions are more variable (0.5128-0.5142), with 9 samples having
 181 $^{143}\text{Nd}/^{144}\text{Nd} > 0.5130$ ($\epsilon_{\text{Nd}} = +10 - +31$). There is no simple correlation between $^{143}\text{Nd}/^{144}\text{Nd}$
 182 and $^{147}\text{Sm}/^{144}\text{Nd}$ (Fig. 3b). In Sr-Nd isotope space the majority of Zealandian xenoliths plot
 183 within, or near to, the field defined by intraplate magmas from New Zealand (Fig. 4a), and
 184 overlap the area that represents the HIMU mantle end-member, as has been previously
 185 observed (Scott et al., 2014a; 2014b). The least metasomatised samples of this study, as for

186 example indicated by $\text{La/Yb}_N < 1$ (Fig. 5), however, preserve significant variability in both Sr
 187 and Nd isotopic compositions expanding the previously observed ranges based on
 188 predominantly metasomatised samples.

189 **3.3. U-Th-Pb isotope data**

190 Measured Pb isotopic compositions are highly variable (e.g. $^{206}\text{Pb}/^{204}\text{Pb} = 17.25-25.42$; Table
 191 2), although the majority of the xenoliths from the Waitaha domain and Chatham Islands
 192 exhibit a more restricted range of $^{206}\text{Pb}/^{204}\text{Pb} = 19.5-20.7$ (Fig. 4). Lead isotopic ratios are
 193 uncorrelated with the abundance of their parent radionuclides (i.e. ^{238}U and ^{232}Th ; Fig. 6a-b),
 194 and show no simple correlations with radiogenic effects in either other lithophile (Nd; Fig. 6c)
 195 or chalcophile (Os; Fig. 6d) isotope systems. The $^{238}\text{U}/^{204}\text{Pb}$ values of the Zealandia xenoliths
 196 range from 0.69 to 552, with 16 of 22 samples from the Waitaha domain having $^{238}\text{U}/^{204}\text{Pb}$
 197 >45 , a value that is significantly higher than in basalts from Mangaia, the most radiogenic
 198 example of classical HIMU OIB localities ($^{238}\text{U}/^{204}\text{Pb} = 26-39$; Woodhead, 1996). In plots of
 199 $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ the data form a linear array that extends from
 200 compositions similar to the classic HIMU mantle component to less radiogenic compositions
 201 (Fig. 4c-d), with the majority of xenoliths having compositions comparable to New Zealand
 202 alkaline intraplate magmas (Fig. 4), which include the carbonatites of the Alpine Dyke Swarm
 203 ($^{206}\text{Pb}/^{204}\text{Pb} = 19.9-20.3$; Barreiro and Cooper, 1987). These basalts show variability based on
 204 their location, with Chatham Island basalts consistently having the highest $^{206}\text{Pb}/^{204}\text{Pb}$ (19.7-
 205 20.8; Panter et al., 2006; Sprung et al., 2007; Timm et al., 2010). Whereas, the xenoliths from
 206 the same location record more variable compositions ($^{206}\text{Pb}/^{204}\text{Pb} = 19.5-22.4$; Table 2), and
 207 elevated $^{207}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$ relative to xenoliths from the Waitaha domain (Fig.
 208 4c). To the west of the Alpine Fault, xenoliths consistently have the lowest Pb isotopic
 209 compositions within Zealandia; in accord with the Pb isotopic compositions of intraplate
 210 basalts especially those from the North Island (e.g. Cook et al., 2005; McGee et al., 2013),

211 which show a greater similarity to MORB compositions with low $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{87}\text{Sr}/^{86}\text{Sr}$,
212 and relatively high ϵ_{Nd} (i.e. $>+6$).

213 **4. DISCUSSION**

214 **4.1. Evidence for multiple Paleoproterozoic to Cenozoic melt depletion events recorded**
215 **by chalcophile and lithophile isotopes**

216 *4.1.1. Re-Os ages*

217 Osmium isotopic compositions are often considered the most robust means of placing age
218 constraints on melting events in the SCLM (e.g. Handler et al., 1997; 2003; Liu et al., 2011;
219 Reisberg and Lorand, 1995; Walker et al., 1989). However, there is continued debate over the
220 meaning of Os model ages in the context of recycling of ancient heterogeneities preserved in
221 the oceanic mantle, perhaps in the form of sulfides or alloys. Liu et al. (2015) report Os
222 isotopic compositions of 14 xenoliths from two Alpine Dyke Swarm localities in West Otago
223 (Fig. 1) that have T_{RD} ages ranging from 0.5-2.1 Ga, with two further xenoliths from Lake
224 Wanaka preserving Archean T_{RD} ages of 2.7 Ga. They suggest that the highly depleted
225 craton-like mantle (forsterite mean = 92.3; $n = 19$; Scott et al., 2014b) underlying West Otago
226 is a mixture of ancient depleted domains and ambient residual mantle that has undergone
227 recent melting prior to being accreted to the SCLM (Liu et al., 2015). This localised
228 occurrence of Archean mantle contrasts with the Os compositions of xenoliths from the
229 Waitaha domain underlying East Otago (McCoy-West et al., 2013), an extensive region
230 ($>55,000 \text{ km}^2$) of more fertile SCLM (Fo mean = 90.5; $n = 27$). The Waitaha domain consists
231 of 7 geographically coherent localities where 16 of 18 samples possess Paleoproterozoic ages,
232 as indicated by a combination of Os model ages, Re-Os isochrons and a regional
233 aluminochron, consistent with widespread melt depletion at 1.9 Ga (McCoy-West et al., 2015;
234 2013). Waitaha domain xenoliths also possess coherent platinum group element patterns that

235 require both a relatively simple melting history and the long-term preservation of their
 236 original mantle sulfides (McCoy-West et al., 2015).

237 *4.1.2. Sm-Nd model ages*

238 The radiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ ($\epsilon\text{Nd} > +30$) of some samples from the Waitaha domain is in
 239 accord with ancient melt depletion. Neodymium model ages were only calculated for samples
 240 that are weakly metasomatised or unmetasomatised ($\text{La/Yb}_\text{N} < 1$), to avoid spurious ages
 241 produced by late stage metasomatism. T_{CHUR} and T_{DM} ages represent maximum and minimum
 242 ages, respectively (Table 1). These samples provide T_{CHUR} model ages varying between 0.1
 243 and 3.5 Ga ($n = 10$), with three samples (and two duplicates) from within the Waitaha domain
 244 recording Nd model ages of 1.5-2.7 Ga (e.g. DPP-3, WFP-2 and WFP-8), consistent with the
 245 regional Paleoproterozoic melting age based on Re-Os systematics (McCoy-West et al., 2015;
 246 2013). Additionally, the steepest array in Sm-Nd isotopic space provides a minimum estimate
 247 of the timing of this ancient melting event at >1525 Ma (Fig. 3; Fig. EA5a). Furthermore,
 248 lithophile evidence for this ancient melting event is observed in the dataset of Scott et al.
 249 (2014b) with two xenoliths having $\epsilon\text{Nd} > +20$, and two different xenoliths having $\epsilon\text{Hf} > 100$,
 250 again consistent with them being residues of Proterozoic melt depletion (>1.5 Ga).

251 Interrogating the record for younger melt depletion events is more complicated.
 252 Several unmetasomatised samples from the Trig L locality have high $^{147}\text{Sm}/^{144}\text{Nd}$ (>0.4; Fig.
 253 3) and produce T_{CHUR} model ages from 130-380 Ma indicative of a secondary Paleozoic-
 254 Mesozoic melting event. The two samples that produce the youngest ages provide a 2-point
 255 age regression of 108 ± 8 Ma (Fig. EA5b). These samples are chemically distinct from the
 256 main cluster of metasomatised samples (Fig. 3) and combined with two additional similar
 257 xenoliths from Trig L in the dataset of Scott et al. (2014b) provide an ‘isochron’ age of $131 \pm$
 258 33 Ma (Fig. EA5b). This is consistent with generation during an Early Cretaceous melting
 259 event (ca. 131-108 Ma). The remaining model ages between 0.5 and 1.5 Ga could be: 1) the

260 result of additional melting events, although it is impossible to distinguish these based on the
 261 lithophile isotope data alone; or 2) mixed ages resulting from the variable contributions of
 262 ancient melting and more recent events (i.e. large amounts of Paleoproterozoic melting and
 263 only weak Cretaceous melting will result in model ages >1 Ga).

264 *4.1.3. Additional Sr-Pb evidence for Paleoproterozoic melt extraction*

265 Strontium isotopic compositions are not widely used for determining ancient melt depletion
 266 events as Rb and Sr compositions of xenoliths are readily overprinted by interaction with high
 267 Rb/Sr crustal fluids containing radiogenic Sr. Although within the Zealandian xenoliths
 268 $^{87}\text{Sr}/^{86}\text{Sr}$ compositions are unsupported by their $^{87}\text{Rb}/^{86}\text{Sr}$, a significant number of samples
 269 still contain unradiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ (<0.7030), with three replicated clinopyroxene separates
 270 preserving $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7019-0.7021, additionally two xenoliths analysed by Scott et al.
 271 (2014b) have $^{87}\text{Sr}/^{86}\text{Sr}$ <0.7022. These are among the least radiogenic Sr isotopic
 272 compositions preserved in xenoliths worldwide. In a similar manner as for Re depletion ages
 273 (e.g. Shirey and Walker, 1998) used in the Re-Os system and for the similar reason that both
 274 Re and Rb are readily overprinted by enriched crustal fluids, it is possible to calculate Rb
 275 depletion model ages. Using a depleted mantle source ($^{87}\text{Rb}/^{86}\text{Sr}$ = 0.0188; $^{87}\text{Sr}/^{86}\text{Sr}$ = 0.7026;
 276 Workman and Hart, 2005) and assuming that all of the Rb was lost during the melting event
 277 (i.e. analogous to a T_{RD} Re-Os age and similarly providing a minimum age of melt depletion).
 278 All 6, low $^{87}\text{Sr}/^{86}\text{Sr}$, samples yield similar Rb depletion model ages of ca. 2.0 Ga (1.91- 2.40
 279 Ga), consistent with the Paleoproterozoic melting event recorded by Os and Nd isotopic data.
 280 Comparable evidence for Proterozoic melt extraction is found in peridotite xenoliths from
 281 Central Asia (Mongolia and Vitim) that preserve similarly unradiogenic Sr isotope
 282 compositions ($^{87}\text{Sr}/^{86}\text{Sr}$ = 0.7017-0.7022) and coupled Sr-Nd and Os model ages of ca. 2 Ga
 283 (Ionov et al., 2005; Pearson et al., 2004; Stosch et al., 1986). Additionally two of the low
 284 $^{87}\text{Sr}/^{86}\text{Sr}$ Zealandia samples have the most unradiogenic Pb isotopic compositions measured

285 ($^{207}\text{Pb}/^{204}\text{Pb} = 15.36\text{-}15.37$); these compositions require ancient melting without subsequent
 286 interaction with a metasomatic agent. Thus only in the rare, unmetasomatised samples do the
 287 Re-Os, U-Pb, Rb-Sr and Sm-Nd systems all record the same Paleoproterozoic melting event.
 288 If this SCLM was the result of recent upwelling and accretion of heterogeneous
 289 asthenospheric mantle to the lithosphere, these lithophile isotope signatures would likely have
 290 been obliterated. Rather, these rare samples provide strong evidence that the widespread
 291 Paleoproterozoic Os model ages from the Waitaha domain xenoliths record a regional melting
 292 event affecting all isotopic systems. For the majority of the xenoliths however, the lithophile
 293 isotope systems have been more heavily overprinted during later metasomatism.

294 **4.2. Determining the signature of the metasomatic agent**

295 Notable positive Th-U and Sr anomalies and large fractionations of Nb/Ta and Ti/Eu in
 296 clinopyroxene have been used to demonstrate that a widespread region of the SCLM of
 297 southern Zealandia has been variably modified by interaction with a carbonatitic component
 298 (McCoy-West et al., 2015; Scott et al., 2014a; 2014b). A strong correlation is observed
 299 between the strength of this metasomatism, as for example reflected in LREE patterns and the
 300 level of isotopic heterogeneity (Fig. 5). The least metasomatised xenoliths ($\text{La}/\text{Yb}_\text{N} < 1$) are
 301 exclusively from the Waitaha domain and preserve highly heterogeneous Sr-Nd-Pb isotopic
 302 compositions ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7019\text{-}0.7044$; $\epsilon_{\text{Nd}} = 4\text{-}32$; $^{206}\text{Pb}/^{204}\text{Pb} = 17.2\text{-}21.3$). Whereas,
 303 strongly metasomatised samples (i.e. Chatham Islands samples with secondary clinopyroxene
 304 and $\text{La}/\text{Yb}_\text{N} = 2.9\text{-}23.2$) possess significantly less isotopic variability and converge to
 305 compositions similar to carbonatites found in the Alpine Dyke Swarm (Barreiro and Cooper,
 306 1987) and pristine oceanic carbonatites (Hoernle et al., 2002; Mourão et al., 2012). Therefore,
 307 the Sr and Nd isotopic composition of the metasomatic agent can be estimated due to the
 308 restricted range of $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} within the metasomatised xenoliths with a composition of

309 $^{87}\text{Sr}/^{86}\text{Sr} = 0.70305 \pm 25$; $\epsilon_{\text{Nd}} = +4.5 \pm 1.5$ (Fig. 5) and in agreement with independent
 310 estimates of Scott et al. (2014b).

311 In contrast, Pb isotopic compositions of even the most metasomatised samples
 312 maintain a significant range (Fig. 5c; $^{206}\text{Pb}/^{204}\text{Pb} = 19.5-21$). The lack of complete
 313 overprinting, as has been the case for Sr and Nd, can be related to the dual lithophile and
 314 chalcophile characteristics of Pb. Sulfides have previously been advocated as the source of the
 315 missing Pb in the silicate mantle (Hart and Gaetani, 2006; Meijer et al., 1990), with included
 316 sulfides in abyssal peridotites preserving direct evidence of long-lived Os and Pb isotopic
 317 heterogeneities in the mantle (Burton et al., 2012). Although no direct estimates of the rate of
 318 Pb diffusion in sulfides are available, Pb is highly compatible in sulfide with estimates of D_{Pb}
 319 sulfide/silicate of ca. 2000 (Burton et al., 2012; Gaetani and Grove, 1999). Additionally, the
 320 low temperatures of the Zealandian SCLM ca. 800-1050 °C (McCoy-West et al., 2015; Scott
 321 et al., 2014b) and of the metasomatic agent (carbonatite magmas erupt at <550 °C; Krafft and
 322 Keller, 1989) will further limit any diffusion. Thus, the Pb isotopic compositions of the
 323 metasomatised xenoliths are considered a mixture of the compositions of the original included
 324 sulfides and the overprinting carbonatite. This provides a mechanism of *in situ* decoupling of
 325 lithophile and chalcophile isotopic signatures with Sr-Nd compositions dominated by the
 326 metasomatic signature and Os dominated by the sulphide and/or metal alloys, whereas Pb
 327 records a mixed signature when mantle sulfides are preserved.

328 **4.3. Age constraints on the widespread metasomatic signature**

329 Previous estimates on the emplacement of this metasomatic signature in the SCLM are highly
 330 variable, but most suggest that the metasomatism occurred during the Phanerozoic between
 331 ca. 500-100 Ma (e.g. Handler et al., 2003; Hart et al., 1997; Panter et al., 2006; 2000; Rocchi
 332 et al., 2002; Scott et al., 2014a; Zhang and O'Reilly, 1997), and prior to the mid-Cretaceous
 333 separation of New Zealand from Antarctica. However, placing strict time constraints on the

334 emplacement of this radiogenic Pb signature in the SCLM is complicated because: 1) it is
 335 unlikely that the xenoliths were in isotopic equilibrium immediately prior to the metasomatic
 336 event; and 2) the preservation of variable amounts of mantle sulfides aids in the development
 337 of Pb isotopic heterogeneities unrelated to the metasomatic agent.

338 *4.3.1. Pb model ages of metasomatism*

339 If we assume the least metasomatised samples in the suite with the lowest $^{238}\text{U}/^{204}\text{Pb}$ and
 340 $^{232}\text{Th}/^{204}\text{Pb}$ values preserve pre-metasomatism isotopic ratios (e.g. OU45852 and WTL-1;
 341 Table 2), and that the current distinctive U-Th/Pb signature was imparted during
 342 metasomatism, we can calculate the length of time required to evolve the isotopic signature
 343 observed in the metasomatised xenoliths (i.e. T_{Meta} ages). Calculated T_{Meta} ages for the
 344 samples, assuming the initial isotopic composition of the region is similar to that of OU45852
 345 ($^{206}\text{Pb}/^{204}\text{Pb} = 19.568$; $^{208}\text{Pb}/^{204}\text{Pb} = 39.056$), using both the ^{238}U - ^{206}Pb and ^{232}Th - ^{208}Pb
 346 systems are generally in good agreement (Table 2; Fig. 7a). These ages are only indicative,
 347 but show radiogenic Pb signatures can evolve very rapidly in the SCLM and that this
 348 metasomatism is a recent event (<180 Ma), with the majority of T_{Meta} age estimates from the
 349 Waitaha domain ≤ 120 Ma ($n = 15/19$; Fig. 7a).

350 *4.3.2. Rapid ingrowth of Pb isotopic signatures*

351 A more robust way of assessing when this metasomatic signature was emplaced into the
 352 SCLM is to look at individual xenolith localities separately, because prior to metasomatism
 353 they probably contained samples closer to isotopic equilibrium. For example, the Chatham
 354 Islands are located ca. 1500 km away from mainland New Zealand and do not record the
 355 Paleoproterozoic melting event observed in the Waitaha domain. Errorchron ages using the
 356 U-Th-Pb chronometers (^{238}U - ^{206}Pb , ^{235}U - ^{207}Pb and ^{232}Th - ^{208}Pb) have been calculated using
 357 whole rock and clinopyroxene data from four localities throughout Zealandia and agree within
 358 error at individual localities (Table 3; Fig. 8). As we can exclude mixing, i.e. there is no

359 known reservoir with the required radiogenic ($^{206}\text{Pb}/^{204}\text{Pb} > 25$) endmember Pb composition,
360 the errorchrons are considered to provide age information, but should be interpreted with
361 caution. All-inclusive ages include all samples from a locality and often have large errors
362 (> 50 Ma) and exceedingly high MSWDs $>> 100$ (Table 3), suggesting they do not represent a
363 single population and a secondary factor has perturbed the isotopic composition of some
364 samples. Therefore, preferred ages have also been calculated from those samples considered
365 to have been in isotopic equilibrium prior to the metasomatism (Fig. 8). The ages discussed
366 below are preferred ages, however, both preferred and all-inclusive ages agree within error
367 and the interpretation outlined is independent of which ages are selected.

368 Whole rock samples from the Chatham Islands produce 4-point U-Th-Pb errorchrons
369 that yield similar ages and are consistent with metasomatism having occurred from ca. 110-
370 120 Ma (e.g. $^{238}\text{U}-^{206}\text{Pb} = 110 \pm 9$ Ma; Table 3; Fig. 8). Further evidence that this
371 metasomatism occurred at ca. 100-120 Ma in the Chatham Islands is provided by the Rb-Sr
372 and Sm-Nd regressions, which provide similar ages (Fig. EA5; Rb-Sr = 117 ± 40 Ma; Sm-Nd
373 = 100 ± 38 Ma), although with large errors due the small range of parent-daughter ratios
374 ($^{87}\text{Rb}/^{86}\text{Sr} = 0.007-0.17$; $^{147}\text{Sm}/^{144}\text{Nd} = 0.12-0.18$). Within the Waitaha domain due to the
375 large range of parent-daughter ratios (e.g. $^{238}\text{U}/^{204}\text{Pb} = 0.7-552$; Fig. 6; Table 2), highly
376 radiogenic Pb isotope ratios can be ingrown vary rapidly. The most radiogenic sample within
377 the xenolith suite WFP-1 ($^{206}\text{Pb}/^{204}\text{Pb} = 25.42$; $^{208}\text{Pb}/^{204}\text{Pb} = 47.28$) also possesses
378 exceptionally elevated $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ at 473 and 2320, respectively. Construction
379 of U-Th-Pb errorchrons for samples from the Fortification Peak locality shows that this
380 composition would require only ca. 67-74 Ma to develop (e.g. $^{232}\text{Th}-^{208}\text{Pb} = 67 \pm 6$; Fig. 8).
381 Xenoliths from the Pilot Point and Trig L localities are consistent with even younger
382 metasomatism with ingrowth of the observed compositions requiring only ca. 28-42 Ma and
383 ca. 24-36 Ma, respectively.

384 The errorchron ages presented here demonstrate that regions of the SCLM with extremely
 385 high U/Pb and Th/Pb as observed in the Zealandian mantle can rapidly evolve to extremely
 386 radiogenic Pb isotopic compositions. We suggest ages from the Chatham Islands are
 387 consistent with this distinctive U-Th-Pb signature being added to the SCLM after ca. 120-110
 388 Ma (Figs. 7 & 8; Fig. EA5), but immediately prior to the cessation of subduction along the
 389 eastern Gondwana margin (e.g. the youngest dated I-type granite in New Zealand is 105 ± 1
 390 Ma; Tulloch and Kimbrough, 2003). Isotopic systematics from three localities within the
 391 Waitaha domain require even younger ages from ca. 70-30 Ma for the ingrowth of the
 392 metasomatic signature. This metasomatism has likely been ongoing with the continuous re-
 393 activation of the fossilised metasomatic signature (Fig. 7b) through interaction with the
 394 sporadic basaltic melts that have traversed the SCLM over the last 100 Ma (McCoy-West et
 395 al., 2010; Timm et al., 2010). In summary, the errorchron ages range from ca. 120 to 30 Ma
 396 and are consistent with this metasomatism being young and likely to have occurred within the
 397 last ca. 120 Ma. Within the current precision, however, it is not possible to resolve definitely
 398 whether the metasomatic signature was added primarily in a distinct event prior to the
 399 cessation of subduction or whether it reflects an average of processes over the last ca. 100 Ma.

400 **4.4. Origins of the radiogenic Pb signature simulating “HIMU” in the lithospheric
 401 mantle**

402 The existence of a widespread HIMU-like signature ($^{206}\text{Pb}/^{204}\text{Pb} > 19.5$) in the southwest
 403 Pacific as observed in intraplate basalts has been extensively investigated (Finn et al., 2005;
 404 Hoernle et al., 2006; McCoy-West et al., 2010; Panter et al., 2006; 2000; Timm et al., 2009;
 405 2010), although whether this component resides in the asthenosphere or SCLM has remained
 406 controversial. The Pb isotopic compositions of mantle xenoliths from throughout Zealandia
 407 preserve similar signatures ($^{206}\text{Pb}/^{204}\text{Pb}$ ca. 19.5-21.5; Fig. 4) allowing us to directly assess the
 408 origin of this “HIMU” signature in the SCLM and whether it is genetically related to a long-

409 lived component in the mantle, i.e. HIMU, or is the result of more recent interaction with
 410 additional fluids or melts. Taking into account the tectonic history of Zealandia, several
 411 processes may have been responsible for the generating this regional “HIMU”-like
 412 component. These are: 1) isolation of an ancient (> 2 Ga) HIMU domain in the SCLM
 413 without secondary modification; 2) overprinting of the xenoliths by the intraplate magmatism
 414 during entrainment and eruption of the xenoliths; 3) metasomatism associated with long-lived
 415 subduction at the eastern margin of Gondwana during the Phanerozoic; and 4) the interaction
 416 of a upwelling mantle with the SCLM immediately prior to and during Gondwana break-up as
 417 proposed in some tectonic models (Hart et al., 1997; Storey et al., 1999; Weaver et al., 1994).

418 *4.4.1. An ancient HIMU component*

419 The generation of the distinctive composition of the HIMU mantle end-member requires a
 420 specific combination of time integrated elevated U/Pb and Th/Pb and also elevated U/Th for
 421 extended periods of time (ca. >0.5 Ga). Although the present day isotopic compositions of
 422 Zealandian xenoliths are consistent with a HIMU-like component (Fig. 4) their Pb isotopic
 423 ratios are unsupported by their parent-daughter ratios (Fig. 6a-b). When comparing the U-
 424 Th/Pb systematics of samples from the Waitaha domain (Table 2; $^{238}\text{U}/^{204}\text{Pb} = 0.7\text{-}552$;
 425 $^{232}\text{Th}/^{204}\text{Pb} = 0.5\text{-}2658$) to the archetypal end-member of classical HIMU (i.e. Mangaia:
 426 $^{238}\text{U}/^{204}\text{Pb} = 32.6 \pm 3.5$; $^{232}\text{Th}/^{204}\text{Pb} = 124 \pm 11$; Woodhead, 1996) the former has significantly
 427 greater ranges, with the variability in the HIMU end-member, assuming an origin from 0.5-
 428 3.0 Ga recycled oceanic crust only increasing to $^{238}\text{U}/^{204}\text{Pb} = 12\text{-}61$ and $^{232}\text{Th}/^{204}\text{Pb} = 45\text{-}177$
 429 (Stracke et al., 2003). A significant number of Waitaha domain samples ($n = 17$) extend to
 430 considerably higher $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ values than required to generate even the
 431 youngest (500 Ma) HIMU signatures (Fig. 6). If this signature resulted from *in situ* ingrowth a
 432 much stronger correlation with isotopic composition would be expected due to the large range
 433 of parent-daughter ratios. Furthermore, there is no correlation between the Pb isotopic

434 compositions and either the Nd or Os isotopic compositions (Fig. 6), which preserve evidence
 435 of Paleoproterozoic melt depletion. Therefore this signature is unrelated to a long-lived
 436 mantle component in the SCLM.

437 *4.4.2. Basalt infiltration*

438 During partial melting of the mantle, lithophile elements are strongly concentrated, by up to
 439 several orders of magnitude, in the melt phase relative to residual peridotite. Thus
 440 contamination of xenoliths during transportation and entrainment within a basaltic melt may
 441 be a possible source of the radiogenic Pb signature. However, several lines of evidence argue
 442 against this: 1) leaching of clinopyroxene separates from the xenoliths has shown that the
 443 most radiogenic Pb is preserved within the crystals and is not the result of grain boundary
 444 contamination (Fig. EA1); 2) the isotopic compositions of the xenoliths and their host
 445 intraplate basalts (Panter et al., 2006; Timm et al., 2009; 2010) are similar (Table EA4) and
 446 therefore irrespective of the higher Pb concentrations in the melts, isotopic mixing will have
 447 little impact; 3) mass balance modelling of trace element budgets has shown that the majority
 448 of xenoliths, can only accommodate very small amounts of basalt infiltration ($\leq 0.1\%$;
 449 McCoy-West et al., 2015) and 4) comparison of Pb metasomatic model ages derived from the
 450 xenoliths with host basalt eruption ages suggests that the metasomatism generally occurred
 451 over an extended period prior to xenolith entrainment (Fig. 7b), rather than at the time of
 452 eruption.

453 *4.4.3. Distinguishing between subduction and plume contributions*

454 Here we propose a model for the origin of the isotopic characteristics of the Zealandian
 455 SCLM incorporating elements of the tectonic history of Zealandia (Fig. 11), although the
 456 exact cause of rifting between Zealandia and Marie Byrd Land remains unclear, having
 457 previously been attributed to a range of processes varying from subducted slab capture
 458 (Luyendyk, 1995) to the impingement of a weak mantle plume on the SCLM (Hart et al.,

459 1997). A long-lived subduction zone was present at the eastern margin of Gondwana,
 460 outboard of the proto-New Zealand from ca. 520-105 Ma (McCoy-West et al., 2014; Muir et
 461 al., 1994, 1996; 1998; Tulloch and Kimbrough, 2003). Invoking this subduction system as the
 462 source of the distinctive Pb isotopic systematics of the SCLM in the southwest Pacific is
 463 straightforward, as has been argued previously for alkaline intraplate magmas (Panter et al.,
 464 2006), however, no current models adequately explain why the locus of rifting did not occur
 465 within the West Antarctic Rift System, the largest continental rift system observed on Earth
 466 that contains extensively thinned continental crust (Rocchi et al., 2002; Winberry and
 467 Anandakrishnan, 2004). Although there are advocates for plume driven rifting, arguments
 468 against a Late Cretaceous plume are based on the lack of large-scale regional uplift
 469 (LeMasurier and Landis, 1996), the much lower than expected magma production rates
 470 relative to common plume-related flood basalt provinces (Finn et al., 2005), and the
 471 complexity of resolving the effects of impingement of an upwelling mantle plume head on the
 472 subducted oceanic lithosphere.

473 Isotopic evidence in mantle xenoliths for subduction being responsible for imparting
 474 this signature into the SCLM is less compelling. During subduction, oceanic crust undergoes
 475 significant dewatering, which lowers the solidus and generates melting in the mantle wedge,
 476 such that any SCLM located above this region for a significant period of time will be
 477 chemically modified by interaction with fluids or melts. Sr isotopic compositions of the
 478 SCLM are easily perturbed by interaction with crustal fluids or melts and in the off-cratonic
 479 mantle are volumetrically dominated by elevated $^{87}\text{Sr}/^{86}\text{Sr} > 0.704$ (e.g. Ionov et al., 2002;
 480 2006; Liu et al., 2012; Witt-Eickschen et al., 2003). Whereas a significant number of
 481 Zealandian xenoliths exhibit unradiogenic $^{87}\text{Sr}/^{86}\text{Sr} < 0.703$; $n = 18$; Table 1), which requires
 482 isolation from fluids derived from altered subducted oceanic crust or the likely overlying
 483 Gondwanan derived sedimentary detritus (i.e. sediments comparable to the Late Mesozoic

484 quartzo-feldspathic greywackes that dominate the Eastern Province: $^{87}\text{Sr}/^{86}\text{Sr} = 0.708\text{-}0.720$;
485 Adams et al., 2002). It is not possible to entirely eliminate that the SCLM has been isolated
486 from subduction zones fluids, rather the fluids may have been generated in an intra-oceanic
487 subduction zone resulting in extremely unradiogenic compositions. Subduction at the eastern
488 margin of Gondwana was long-lived (ca. 520-105 Ma) and uniformitarianism would suggest
489 that if this process was ongoing for >100 m.y. a much more widespread and regionally
490 homogeneous distribution of this radiogenic Pb signature would be expected along the paleo-
491 subduction margin, however, in eastern Australia this signature is not observed (i.e.
492 $^{206}\text{Pb}/^{204}\text{Pb} < 19.5$; Nasir et al., 2010; Paul et al., 2005; Zhang et al., 2001). Instead, tectonic
493 reconstructions prior to Gondwana break-up show a somewhat concentric distribution of this
494 HIMU-like Pb composition centred around the boundary between the Campbell Plateau and
495 Marie Byrd Land (i.e. an elongate bullseye: see Fig. 8 in Finn et al., 2005), consistent with the
496 impingement of upwelling mantle on the SCLM, that was constrained within the mantle
497 wedge and therefore spread parallel to the trench axis. Additionally, due to the large range in
498 $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ values within the Zealandian mantle xenoliths, which can evolve to
499 very distinctive compositions in <100 m.y. (see 5.3), significantly more Pb isotopic
500 heterogeneity would be expected if subduction was responsible for imparting this
501 metasomatic signature over an extended period of the Phanerozoic (>400 m.y.).
502 Furthermore, the distinctive clinopyroxene trace element patterns (McCoy-West et al., 2015;
503 Scott et al., 2014b) observed within Zealandian mantle xenoliths are incompatible with
504 generation by subduction zone fluids (i.e. lack of fractionation between U and Th) and have
505 instead been attributed to carbonatite metasomatism (McCoy-West et al., 2015; Scott et al.,
506 2014a; 2014b). Isotopic evidence is also consistent with a carbonatite component as: 1) the
507 main cluster of strongly metasomatised xenoliths have similar isotopic compositions as
508 measured in unmodified oceanic carbonatites (Fig. 5); and 2) the unradiogenic $^{87}\text{Sr}/^{86}\text{Sr}$

509 observed in Waitaha xenoliths requires that the metasomatic agent is low in Rb, another
 510 common feature of carbonatites (Hoernle et al., 2002; Ray et al., 2000). Carbonatite
 511 metasomatism is highly effective at transporting large ion lithophile elements such as U
 512 (Green and Wallace, 1988), and in some weakly metasomatised samples small volume
 513 fractionations have produced extreme $^{238}\text{U}/^{204}\text{Pb}$ ratios with the samples rapidly developing
 514 negative $\Delta 7/4$ (Fig. 9; $\Delta 207\text{Pb}/204\text{Pb}$; Hart, 1984) as they evolve along a trajectory away
 515 from the mantle array (see Fig. 10). Samples with the most negative $\Delta 7/4$ also possess
 516 strongly fractionated Nb/Ta in their clinopyroxene (Fig. 9), as would be expected from
 517 disequilibrium interactions with a carbonatitic agent (Green et al., 1992). Carbonatites
 518 erupted within the Alpine Dyke Swarm could possibly represent the carbonatitic metasomatic
 519 agent, however, it is not possible to rigorously assess this due to the lack of coupled U-Th and
 520 Pb isotopic data (Barreiro and Cooper, 1987; Cooper, 1986; Cooper and Paterson, 2008) for
 521 this suite.

522 Although controversial, the involvement of upwelling mantle, possibly in the form of a
 523 plume feature, has been suggested to have been involved in the separation of Zealandia from
 524 West Antarctica (Hart et al., 1997; Kipf et al., 2014; Lanyon et al., 1993; Storey et al., 1999;
 525 Sutherland et al., 2010; Weaver et al., 1994). Weaver *et al.* (1994) placed the axis of their
 526 proposed plume at the reconstructed boundary between the Campbell Plateau and Marie Byrd
 527 Land, intersecting the region defined by the most radiogenic Pb values ($^{206}\text{Pb}/^{204}\text{Pb} > 20.5$)
 528 prior to Gondwana break-up (Finn et al., 2005; Panter et al., 2006) and adjacent to the oldest
 529 oceanic crust observed between New Zealand and Antarctica (Laird and Bradshaw, 2004).
 530 The distinctive Pb isotopic signature of the Zealandian xenoliths is consistent with addition of
 531 this metasomatic signature to the SCLM in the Late Cretaceous sometime after ca. 120 Ma
 532 (Fig. 8). Additionally, elevated $^{147}\text{Sm}/^{144}\text{Nd}$ and Nd isotopic compositions of three samples
 533 within the Waitaha domain, which are distinct from the composition of the metasomatic

534 agent, suggest a secondary melting event in the Early Cretaceous (ca. 131-108 Ma; Fig.
535 EA5b). Disturbance of ancient Nd model ages in least metasomatised samples from the
536 Waitaha domain is compatible with this melting event being widespread. During the Early
537 Cretaceous the Pacific Plate reversed polarity and started moving slowly northwards (i.e.
538 Aptian: 125-112 Ma; Bradshaw, 1989; Larson et al., 1992), with collision between the
539 Hikurangi Plateau and the northern margin of Zealandia resulting in a cessation of subduction
540 at ca. 105 Ma along this portion of the Gondwana margin (Davy et al., 2008; Hoernle et al.,
541 2010; Mukasa and Dalziel, 2000). Extensional volcanism and major fault-bounded
542 sedimentary basins developed shortly thereafter (i.e. Stitts Tuff: 101 ± 2 Ma; Muir et al.,
543 1997), during a period of major crustal extension prior to the opening of the Tasman Sea. We
544 suggest a combination of these processes weakened the subducting lithosphere and
545 subsequently resulted in detachment of the slab allowing a pathway for upwelling mantle to
546 interact with the SCLM (Fig. 11b). The impingement of this upwelling mantle on the
547 lithosphere would have resulted in elevated heat flow, as recorded by the voluminous
548 Separation Point Suite ($>20,000 \text{ km}^2$) that was emplaced into the crust between ca. 120-110
549 Ma (Mortimer et al., 1997; Muir et al., 1994; 1997).

550 Therefore, on the basis of tectonic and isotopic evidence we propose that the most
551 likely scenario is the radiogenic HIMU-like Pb signature was emplaced by carbonatite-rich
552 low degree melts in the periphery of a large weak plume head that metasomatised the SCLM
553 sometime after 120 Ma. This upwelling mantle either punched through a weakened subducted
554 slab or was the catalyst for its detachment, although subduction had entirely ceased by ca. 105
555 Ma. During upwelling these carbonatite-rich melts variably metasomatised a widespread
556 region of SCLM (ca. 5 million km^2 ; Fig. 11b). This metasomatism is possibly ongoing to the
557 present day with the “fossilised” carbonatite signature remobilised when asthenospheric melts
558 traverse the SCLM producing the distinctive intraplate magmas of Zealandia (Fig. 11d).

559 **4.5. Rapid ingrowth of signatures simulating “HIMU” and more heterogeneity in the
560 future**

561 The extreme range of $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ observed in Zealandian xenoliths
562 unsupported by their Pb isotopic compositions (Fig. 6), requires the recent addition of this U-
563 Th enriched component to the SCLM. If undisturbed ingrowth of radiogenic Pb continues, a
564 previously unrecognised signature with extreme Pb isotopic compositions (“ultra HIMU”)
565 will rapidly develop. Forward modelling of xenolith Pb isotopic compositions, using their
566 present day parent-daughter ratios is presented (Fig. 10; Table EA5). In the future the Waitaha
567 domain will possess a significantly more heterogeneous composition (i.e. in 200 m.y.
568 $^{206}\text{Pb}/^{204}\text{Pb} = 17.5\text{--}40.3$), especially in $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic space, due to the
569 variable Th/U of the samples. For comparison, *sensu stricto* HIMU samples from Mangaia
570 evolve as a coherent group and preserve higher $^{207}\text{Pb}/^{204}\text{Pb}$ (Fig. 10a), but after only ca. 150
571 m.y. some samples from the heterogeneous Waitaha domain will have both higher $^{206}\text{Pb}/^{204}\text{Pb}$
572 and $^{208}\text{Pb}/^{204}\text{Pb}$ (Fig. 10b). Outside of Zealandia, radiogenic Pb signatures in the off-cratonic
573 mantle are generally rare, see the xenolith compilation of Wittig et al. (2009; $n = 350$).
574 HIMU-like compositions attributed to carbonatite metasomatism are also observed in
575 peridotite xenoliths from the Atlas Mountains, Morocco (Marks et al., 2009; Wittig et al.,
576 2010). The Atlas xenoliths have a restricted range of Pb isotope compositions ($^{206}\text{Pb}/^{204}\text{Pb} =$
577 19.98–20.25) and comparable $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ to the main cluster of Zealandian
578 xenoliths (Fig. 6). Forward modelling done by Wittig et al. (2010) results in less radiogenic
579 compositions but a similar trajectory in $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ space, although the Atlas
580 samples evolve more rapidly in $^{208}\text{Pb}/^{204}\text{Pb}$ space due to their higher Th/U (Fig. 10).
581 Additionally, a few studies have attributed the development of HIMU-like signatures in
582 continental intraplate basalts to radiogenic ingrowth within a carbonate metasomatised SCLM
583 (Janney et al., 2002; Tappe et al., 2007). The fact that this signature is not more widespread is

584 surprising, but could be attributed to: 1) the fortuitous rapid removal of this signature from the
 585 SCLM during subsequent melting events effectively removing the excess U and Th; or 2) the
 586 process of generating large fractionations in Th/Pb and U/Pb by carbonatite related
 587 metasomatism has only become widespread on the Phanerozoic Earth (64% of dated
 588 carbonatites are Phanerozoic in age, although this could be a preservation bias; Veizer et al.,
 589 1992; Wolley and Bailey, 2012), due to secular changes in the oxygen fugacity of the mantle
 590 (Evans, 2012; Shirey and Richardson, 2011; Tappe et al., 2014) that have increased
 591 carbonatite stability during interactions at the base of the SCLM.

592 Although the exact mechanism that imparted the radiogenic Pb signature into the
 593 SCLM in Zealandia may be debated, it has been demonstrated that this signature is young and
 594 has no genetic link with the classical HIMU mantle end-member observed in some OIB
 595 defined by Zindler & Hart (1986). Due to the antiquity of much of the sampled SCLM, its
 596 isotopic composition results from multiple melting and metasomatic events. Therefore,
 597 classifying samples from the SCLM using nomenclature originally intended for oceanic
 598 basalts has misleading connotations regarding the longevity of these signatures (i.e.
 599 heterogeneity within OIB and MORB is the result of the segregation of distinct components
 600 and 1-3 G.y. of storage in the deep mantle). Studies worldwide describe the presence of
 601 SCLM with HIMU affinities, largely based on the compositions of intraplate basalts (e.g.
 602 Panter et al., 2006; Rooney et al., 2014), although, whether any of these samples really
 603 contain an archetypal HIMU or are instead the result of recent interaction with a metasomatic
 604 agent remains unresolved. We suggest that the process recorded here whereby carbonatite
 605 metasomatism imparts extremely elevated parent-daughter ratios ($^{238}\text{U}/^{204}\text{Pb} > 70$; $^{232}\text{Th}/^{204}\text{Pb}$
 606 > 200) to regions of the SCLM be henceforth distinguished as CarboHIMU (i.e. carbonatite
 607 related high μ).

608

609 **5. CONCLUSIONS**

610 An integrated U-Th-Pb, Rb-Sr, Sm-Nd and Re-Os isotopic dataset for spinel peridotite
611 xenoliths sampling the variably metasomatised SCLM beneath Zealandia, allows new insights
612 into the formation of off-cratonic lithosphere and time constraints to be placed on the
613 development of the “HIMU”-like signature observed throughout the southwest Pacific. Rare
614 weakly metasomatised xenoliths preserve the coupling of chalcophile (Os) and lithophile (Sr
615 and Nd) isotope compositions, all indicative of melt extraction at ca. 2.0 Ga. Furthermore,
616 Paleoproterozoic Re-Os model ages are regionally widespread in the more abundant, strongly
617 metasomatised xenoliths, with the lithophile isotopic signatures having been subsequently
618 overprinted by carbonatite-rich melts. These observations are consistent with the formation
619 and stabilisation of a voluminous region of SCLM (≥ 2 million km³) in a Paleoproterozoic
620 melt depletion event and the preservation of this ancient mantle keel beneath a large portion
621 of Zealandia today. This contrasts with interpretations from a localised occurrence of
622 xenoliths with ancient (up to 2.7 Ga) Os model ages, decoupled from other isotopic systems,
623 as mantle residues recycled from within the convecting mantle and later accreted to the SCLM
624 (Liu et al., 2015). Lead isotope compositions are decoupled from the other isotopic data, with
625 extreme $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ values requiring this metasomatic signature to have been
626 added to the SCLM in the Early Cretaceous and continuously reactivated by subsequent
627 volcanism. We propose this distinctive radiogenic Pb signature was imparted into the SCLM
628 by carbonate-rich low degree melts that heterogeneously metasomatised a widespread region
629 of SCLM sometime after ca. 120 Ma, but prior to the onset of this type of volcanism at 98 Ma.
630 If allowed to develop undisturbed into the future, the extreme U/Pb and Th/Pb characteristics
631 of these samples will produce extreme Pb isotopic values distinct from any reservoir currently
632 observed on Earth. We suggest the limited sampling of this signature in the rock record may
633 reflect a secular change in mantle chemistry associated with the increased prevalence of

634 carbonatite sources within the mantle and that this unique metasomatic signature be referred
635 to as CarboHIMU.

636

637

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915 **Figure Captions**

916 *Figure 1:* Simplified geological map of New Zealand's crustal basement showing the mantle
 917 xenolith localities. Crustal geology is based on Mortimer (2004). Mantle xenoliths are divided
 918 into three groups: (1) Waitaha domain (circles); Chatham Islands (squares); and (3) all other
 919 localities (diamonds) based on Re-Os isotopic characteristics (McCoy-West *et al.*, 2013). The
 920 dashed ellipse represents the minimum extent of the Paleoproterozoic Waitaha domain of
 921 lithospheric mantle. Previously studied xenolith localities are shown divided as follows; those
 922 considered to be within the Waitaha domain (black circles; Scott *et al.*, 2014b), the Auckland
 923 Islands (blue square) and West Otago (grey diamonds; Liu *et al.*, 2015).

924

925 *Figure 2:* Comparison of concentrations and ratios of trace elements in unleached
 926 clinopyroxene measured by laser ablation (McCoy-West *et al.*, 2015) and clinopyroxene
 927 residues after leaching analysed by TIMS. (a-b) concentrations of Sr and Pb, respectively. (c-
 928 f) Sm/Nd, Th/U, U/Pb and Th/Pb ratios, respectively. Samples with any element with a
 929 concentration measured by laser of <15 ppb are shown white. The leachates of the high
 930 concentration sample DPP-1 as measured by TIMS have isotopic ratios (e.g. Th/U) far from
 931 unity, whereas the leached residue of this sample plots on the 1:1 line.

932

933 *Figure 3:* Comparison of (a) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ and (b) $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$
 934 isotopic systematics of mantle xenoliths from around Zealandia. Mantle xenoliths are divided
 935 into three groups: (1) Waitaha domain (circles); Chatham Islands (squares); and (3) all other
 936 localities (triangles) based on Re-Os isotopic characteristics (McCoy-West *et al.*, 2013). (a)
 937 The Waitaha domain exhibits greater isotopic heterogeneity, with no simple correlation
 938 observed between $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$. (b) Several arrays appear to form in Sm-Nd isotopic
 939 space, consistent with at least two events; Paleoproterozoic and Cretaceous melting with the
 940 remainder probably being mixed ages. Comparative xenolith data (Scott *et al.*, 2014a; 2014b)
 941 are separated into samples considered to be within the Waitaha domain (North and East
 942 Otago; black circles; n = 28), and those from the Auckland Islands (blue squares; n = 4).

943

944 *Figure 4:* Strontium-Nd-Pb isotopic plots comparing Zealandian mantle xenoliths to oceanic
 945 mantle end-member compositions and intraplate basalts from the New Zealand region. Whole
 946 rock analyses (triangles) and clinopyroxene separates (squares) are distinguished. Error bars
 947 are generally significantly smaller than the size of the symbols, except where shown Pb

948 analyses. Data sources for mantle end-member compositions are provided in Fig. EA4. The
 949 archetypal example of HIMU end-member from the Cook-Austral Islands, Mangaia is shown
 950 separately (Woodhead, 1996). New Zealand intraplate basalts have been divided into two
 951 groups on the basis of their location, the North Island (Cook et al., 2005; Huang et al., 2000;
 952 Huang et al., 1997; McGee et al., 2013; Sprung et al., 2007; Timm et al., 2010) and southern
 953 Zealandia (i.e. the South Island and all the outlying Islands on the Campbell Plateau; Barreiro
 954 and Cooper, 1987; Hoernle et al., 2006; McCoy-West et al., 2010; Panter et al., 2006; Sprung
 955 et al., 2007; Timm et al., 2009; Timm et al., 2010), with samples showing evidence of crustal
 956 contamination excluded (i.e. consistent with assimilation of Torlesse Greywacke: $\text{SiO}_2 > 52$ wt
 957 % or $^{87}\text{Sr}/^{86}\text{Sr} > 0.7038$; McCoy-West et al., 2010). NRHL = northern hemisphere reference
 958 line: Hart (1984). Mantle xenoliths are compared to mantle components uncorrected for the
 959 host basalt eruption age, this correction makes negligible difference to the majority of Sr-Nd
 960 data, and has minimal effect on Pb data for the majority of samples due to the young eruption
 961 ages (<86 Ma), and is only significant for those samples with extremely elevated $^{238}\text{U}/^{204}\text{Pb}$
 962 >100 (see Fig. EA4).

963

964 *Figure 5:* Lithophile isotope (Sr-Nd and Pb) variation of Zealandian mantle xenoliths relative
 965 to their degree of metasomatism (i.e. whole rock La/Yb_N ; N = chondrite-normalised; Palme &
 966 O'Neill, 2014). Unmetasomatised samples ($\text{La}/\text{Yb}_N < 1$) from the Waitaha domain possess the
 967 greatest isotopic heterogeneity. As the degree of metasomatism increases Sr and Nd isotopic
 968 compositions converge to that of the metasomatic agent (e.g. grey bar; $^{87}\text{Sr}/^{86}\text{Sr} = 0.7028$ -
 969 0.7033; $\varepsilon_{\text{Nd}} = +3$ - $+6$), while Pb isotopic compositions remain more heterogeneous ($^{206}\text{Pb}/^{204}\text{Pb}$
 970 = 19.5-21). Comparative xenolith data (Scott et al., 2014a; 2014b) has been plotted using the
 971 La/Yb_N of clinopyroxene. Also shown are Alpine Dyke Swarm carbonatites ($\text{La}/\text{Yb}_N >> 25$;
 972 Barreiro and Cooper, 1987) and oceanic carbonatites unaffected by crustal contamination
 973 from Fogo and Brava, Cape Verde Islands (Hoernle et al., 2002; Mourão et al., 2012).

974

975 *Figure 6:* Comparison of Pb isotopic compositions of Zealandian mantle xenoliths with other
 976 isotopic data. The shaded bar represents a restricted range of Pb isotopic compositions
 977 ($^{206}\text{Pb}/^{204}\text{Pb} = 19.5$ - 20.5 ; $^{208}\text{Pb}/^{204}\text{Pb} = 39$ - 40) to illustrate the lack of correlation between Pb
 978 isotopic ratios and parent-daughter ratios (a-b) or other isotopic systems (Nd and Os; c-d). (a-
 979 b) the dotted lines show the range of $^{238}\text{U}/^{204}\text{Pb}$ and $^{232}\text{Th}/^{204}\text{Pb}$ required to explain the
 980 variability observed in the HIMU end-member based on 0.5-3.0 Ga old recycled oceanic crust

981 ($^{238}\text{U}/^{204}\text{Pb} = 12\text{-}61$; $^{232}\text{Th}/^{204}\text{Pb} = 45\text{-}177$; Stracke et al., 2003). Mangaia represents the
 982 archetypal example of the classical HIMU end-member (Woodhead, 1996), with peridotite
 983 xenoliths from the Atlas Mountains, Morocco also shown (Wittig et al., 2010). (d) the
 984 $^{187}\text{Os}/^{186}\text{Os}$ isotopic composition of the PUM (primitive upper mantle) is 0.1296 ± 8 (Meisel
 985 et al., 2001) with the age scale showing the approximate Re-depletion model ages based on
 986 these compositions.

987

988 *Figure 7:* (a) Histogram showing average ^{206}Pb and ^{208}Pb T_{Meta} metasomatic model ages for
 989 xenoliths from the Waitaha domain and Chatham Islands. Relative probability curves,
 990 assuming a 20 Ma error on the ages, are shown for ^{238}U - ^{206}Pb (solid line) and ^{232}Th - ^{208}Pb
 991 (dotted line) metasomatic ages (see Table 2). (b) Plot of eruption age of the host basalt versus
 992 the average metasomatic model ages of mantle xenoliths. Also shown are the errorchron ages
 993 for individual localities (see Table 3). The filled box plot represents the weighted average of
 994 the preferred errorchrons and a 95% confidence interval error, with the dotted line
 995 representing the all-inclusive 95% confidence interval error. Ages of geologic events are
 996 taken from Eagles et al. (2004); Hoernle et al. (2010); Kamp (1986); McCoy-West et al.
 997 (2010).

998

999 *Figure 8:* ^{238}U - ^{206}Pb and ^{232}Th - ^{208}Pb errorchrons of mantle xenoliths from individual localities
 1000 within Zealandia, the Chatham Islands (a-b), Fortification Peak (c-d), Pilot Point (e-f) and
 1001 Trig L (g-h). These ages are only considered indicative of the timing of metasomatism and
 1002 assume the samples were initially in isotopic equilibrium. However, they demonstrate how
 1003 quickly ingrowth of these radiogenic Pb signatures can occur with such extreme parent-
 1004 daughter ratios. Errorchron ages are preferred ages (solid lines; see Table 3). Samples
 1005 plotted white are excluded from the age calculations with the grey ages and dotted lines
 1006 showing the effect of including these samples.

1007

1008 *Figure 9:* Unique Pb isotopic characteristics of Zealandian xenoliths due to metasomatism: (a)
 1009 $^{238}\text{U}/^{204}\text{Pb}$ versus $\Delta 7/4$; and (b) Nb/Ta of clinopyroxene crystals versus $\Delta 7/4$. Hollow
 1010 symbols represent individual crystal analyses, with the filled symbols showing the average
 1011 value for the xenolith. $\Delta 7/4$ is the vertical deviation in $^{207}\text{Pb}/^{204}\text{Pb}$ calculated relative to the
 1012 northern hemisphere reference line (Hart, 1984). The majority of samples are strongly
 1013 metasomatised and have Pb isotopic compositions similar to Cape Verde carbonatites

1014 (Hoernle et al., 2002; Mourão et al., 2012) and Atlas mountains xenoliths (Wittig et al., 2009;
 1015 Wittig et al., 2010) whereas, weakly metasomatised samples have not equilibrated with the
 1016 metasomatic agent and preserve strongly negative $\Delta7/4$ coupled with extreme $^{238}\text{U}/^{204}\text{Pb}$ and
 1017 Nb/Ta fractionations consistent with carbonatite metasomatism.

1018

1019 *Figure 10:* Modelled Pb isotopic compositions of mantle xenoliths from the Waitaha domain
 1020 200 m.y. in the future. White symbols represent the present day measured compositions; with
 1021 filled colour symbols representing the modelled isotopic compositions (see Table EA5). The
 1022 blue circles joined by dotted lines represent the evolution of three distinct samples in 50 m.y.
 1023 intervals. The five samples with the most extreme $^{238}\text{U}/^{204}\text{Pb}$ already plot significantly off the
 1024 graph (see Table EA5). The evolution of Mangaia, the archetypal example of the HIMU end-
 1025 member (Woodhead, 1996), and Atlas Mountain xenoliths (Wittig et al., 2010), over the same
 1026 time interval are shown for comparison. The variability of U/Pb, Th/Pb and Th/U within the
 1027 xenoliths will result in the rapid development of highly radiogenic and heterogeneous Pb
 1028 isotope signature in the future, with compositions different from any known mantle reservoirs.

1029

1030 *Figure 11:* Schematic diagram showing the complex history of the lithospheric mantle under
 1031 Zealandia during the Phanerozoic. The nascent lithosphere of the Waitaha domain formed at
 1032 ca. 1.9 Ga on the margin of the Paleoproterozoic supercontinent Nuna, subsequently the
 1033 Waitaha domain SCLM became partially or totally decoupled from its complementary
 1034 overlying crust and remained isolated from other major crustal bodies without being
 1035 significantly reworked for >1 G.y. (McCoy-West et al., 2013). (a) The Waitaha domain
 1036 arrived at the eastern margin of Gondwana after ca. 280 Ma (Kimbrough et al., 1992), because
 1037 of its present day position to the east of the Dun Mountain Ophiolite Belt, preserving evidence
 1038 of a major oceanic basin closure event. (b) Carbonate-rich low degree melts in a weak mantle
 1039 upwelling, possibly in the form of a mantle plume, heterogeneously metasomatised the SCLM
 1040 imparting a distinctive elevated U/Pb and Th/Pb signature (CarboHIMU). This event is
 1041 interpreted to occur in the Early Cretaceous, ca. 120-110 Ma, based on melting and
 1042 metasomatic ages (see Figs. 7; 8; EA5). (c) Sea-floor spreading between Zealandia and Marie
 1043 Byrd Land, West Antarctica, is constrained at ca. 84 Ma by the oldest oceanic crust adjacent
 1044 to the Campbell Plateau (Eagles et al., 2004). (d) Over the last 84 Ma alkaline intraplate
 1045 volcanism has been ongoing while Zealandia has drifted 6000 km northwest (Timm et al.,
 1046 2010). These melts have continuously reactivated the metasomatic signature producing the

1047 young metasomatic ages observed (Fig. 8). Arrows represent the direction and relative
1048 strength of mantle flow.

1049

1050

1051

1052 **Table Footnotes**1053 *Table 1:*

1054 ^ASample type: ¹ whole rock powder; ² clinopyroxene separate; ³ orthopyroxene separate. # is a
 1055 replicate digestion of the sample. Replicate digestions of sample FvF-1 contained ^c coarse
 1056 (425-500 µm) and ^f fine (275-425 µm) clinopyroxene.

1057 ^BWhole rock Al₂O₃ contents from McCoy-West et al. (2015).

1058 ^CWhole rock La/Yb_N (chondrite-normalised; Palme and O'Neill, 2014) ratios are calculated
 1059 from data in McCoy-West et al. (2015). *For samples from the Pilot Point locality
 1060 clinopyroxene ratios were used as whole rock compositions are disturbed by alteration
 1061 (McCoy-West et al., 2015).

1062 ^DInitial ¹⁸⁷Os/¹⁸⁸Os is taken from McCoy-West et al. (2013).

1063 ^EInitial isotopic compositions are calculated using the age of eruption of the host basalt at
 1064 each locality, which can be found in (McCoy-West et al., 2015).

1065 ^F $\epsilon_{\text{Nd}}(T) = ((^{143}\text{Nd}/^{144}\text{Nd}_{\text{sam}}(T) / ^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}}(T)) - 1) \times 10^4$ where T is the age of eruption
 1066 of the host basalt and is calculated using the CHUR compositions below.

1067 ^GNd model ages are only shown for those samples that have been weakly metasomatised (i.e.
 1068 La/Yb_N < 1). T_{CHUR} was calculated assuming ¹⁴⁷Sm/¹⁴⁴Nd = 0.1967 and ¹⁴³Nd/¹⁴⁴Nd =
 1069 0.512638 (Jacobsen and Wasserburg, 1984). T_{DM} was calculated assuming ¹⁴⁷Sm/¹⁴⁴Nd =
 1070 0.2136 and ¹⁴³Nd/¹⁴⁴Nd = 0.51315, the depleted MORB mantle composition of Workman and
 1071 Hart (2005).

1072 Sr, Sm and Nd concentration data are precise to better than 0.02 %, 0.05 % and 0.05 %,
 1073 respectively. Rb concentrations are generally precise to 0.2-0.4 %, although rare samples with
 1074 the lowest concentrations have up to a 2.5 % error.

1075

1076 *Table 2:*

1077 ^ASample type: ¹ whole rock powder; ² clinopyroxene separate; ³ orthopyroxene separate. # is a
 1078 replicate digestion of the sample. Replicate digestions of sample FvF-1 contained ^c coarse
 1079 (425-500 µm) and ^f fine (275-425 µm) clinopyroxene.

1080 ^BInitial Pb isotopic compositions are calculated using the age of eruption of the host basalt at
 1081 each locality, which can be found in McCoy-West et al. (2015).

1082 ^CΔ7/4 and Δ8/4 are the vertical deviation in ²⁰⁷Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb calculated relative to
 1083 the northern hemisphere reference line (Hart, 1984).

1084 ^DT_{Meta}²⁰⁶ and T_{Meta}²⁰⁸ provide an estimate of how long it would take to evolve the measured
 1085 ²⁰⁶Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb ratios, respectively, in the samples assuming that metasomatism
 1086 created their elevated U/Pb and Th/Pb ratios and their original isotopic composition were that
 1087 of unmetasomatised sample OU45852 (²⁰⁶Pb/²⁰⁴Pb = 19.568; ²⁰⁸Pb/²⁰⁴Pb = 39.056).

1088 U, Th and Pb concentration data are generally precise to better than 0.1 %, 0.3 % and 0.001 %,
 1089 respectively.

1090

1091 *Table 3:*

1092 Errorchron age estimates were calculated using isoplot (Ludwig, 2008), using either the
 1093 measured errors or a 1% error on the parent daughter ratio (whichever was larger), and the
 1094 reproducibility of Pb standard isotope measurements over a 2 year period (2σ population:
 1095 ²⁰⁶Pb/²⁰⁴Pb = ± 0.018; ²⁰⁷Pb/²⁰⁴Pb = ± 0.013; ²⁰⁸Pb/²⁰⁴Pb = ± 0.055). Weighted averages are
 1096 calculated using the three linked decay schemes.

1097 ^AThe assumption underpinning the errorchrons is that the samples were in isotopic
 1098 equilibrium prior to the addition of the metasomatic agent. All-inclusive ages are based on
 1099 data from all samples analysed for a given locality. Preferred ages exclude any samples were
 1100 the assumption of prior equilibrium clearly does not hold (i.e. data points are greater than 3x

1101 the external Pb reproducibility, e.g. = ± 0.054 (2σ) for $^{206}\text{Pb}/^{204}\text{Pb}$, from the regression formed
1102 by the other points).

1103 ^B n = the number of samples included in a calculation. Numbers in parentheses represent the
1104 total number of xenoliths analysed at a locality.

1105 ^C Reasons for exclusion: Chatham Islands; Clinopyroxene separates were the smallest
1106 measured in this study (<10 mg) and preserve heterogeneity unrelated to the metasomatism.
1107 Fortification Peak; Sample WFP-8 falls below the regression line consistent with being a
1108 Paleoproterozoic melt residue ($\varepsilon_{\text{Nd}} > +27$) that preserves an unradiogenic Pb composition that
1109 has not equilibrated with the metasomatic agent. Pilot Point; the small whole rock xenoliths
1110 do not preserve evidence of the unradiogenic Pb composition seen in the clinopyroxene
1111 separates. MSWD $\gg 1000$ suggests that this is not a single population. Trig L; Sample
1112 WTL-3 is excluded due to its significant offset in $^{206}\text{Pb}/^{204}\text{Pb}$ from the other xenoliths in this
1113 overall weakly metasomatised locality (McCoy-West et al., 2015).

1114

Table 1: Rb-Sr and Sm-Nd concentration and isotopic data for whole rocks and mineral separates from New Zealand mantle xenoliths.

Sample ^A	Al ₂ O ₃ ^B (wt %)	La/ Yb _N ^C	¹⁸⁷ Os/ ¹⁸⁸ Os _(i) ^D	Rb (ppm)	Sr (ppm)	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr ^E measured	⁸⁷ Sr/ ⁸⁶ Sr ^E initial	Sm (ppm)	Nd (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd ^E measured	¹⁴³ Nd/ ¹⁴⁴ Nd ^E initial	ϵ_{Nd} ^F	T _{CHUR} ^G (Ga)	T _{DM} ^G (Ga)
North Island																
NGB-3 ¹	0.06	2.76	0.12700	0.0347	4.583	0.0219	0.703532 ± 12	0.70353	0.0200	0.0072	0.2163	0.513029 ± 12	0.51303	7.6 ± 2	--	--
NGB-4 ¹	0.16	29.8	0.12942	0.1159	4.691	0.0714	0.703183 ± 9	0.70318	0.0881	0.0169	0.1158	0.512953 ± 7	0.51295	6.2 ± 1	--	--
South Island																
P43153b ¹	1.02	9.48	0.11242	0.5356	3.022	0.5128	0.709371 ± 5	0.70878	0.0987	0.0224	0.1373	0.512895 ± 5	0.51282	5.0 ± 1	--	--
P45280 ¹	0.53	0.873	0.12584	0.0554	0.383	0.4184	0.703340 ± 8	0.70282	0.0534	0.0197	0.2228	0.512952 ± 8	0.51282	6.1 ± 2	1.83	-3.34
MSI33D ¹	0.71	8.84	0.12406	0.6785	6.696	0.2930	0.703061 ± 5	0.70295	0.5962	0.1709	0.1733	0.512869 ± 3	0.51284	4.5 ± 1	--	--
MSIK33C ¹	1.73	4.81	0.12630	0.0648	2.789	0.0672	0.703448 ± 8	0.70342	0.3641	0.1209	0.2008	0.512872 ± 8	0.51284	4.6 ± 2	--	--
<i>Waitaha domain</i>																
MSI20C ²	1.26	16.54	0.11982	0.0193	67.84	0.0008	0.702850 ± 5	0.70285	2.165	0.2869	0.0801	0.512950 ± 6	0.51295	6.1 ± 1	--	--
MSI20G ¹	1.62	15.22	0.11989	0.0228	2.768	0.0238	0.702966 ± 7	0.70296	0.1123	0.0190	0.1022	0.512926 ± 16	0.51292	5.6 ± 3	--	--
MSI20G ^{1#}									0.0940	0.0155	0.1000	0.512928 ± 12	0.51292	5.7 ± 2	--	--
FvF-1 ^{2c}	2.61	0.888		0.3471	33.41	0.0300	0.702645 ± 9	0.70264	1.649	0.6333	0.2323	0.513469 ± 8	0.51342	16.2 ± 2	3.53	2.59
FvF-1 ^{2f}				0.4536	36.31	0.0361	0.702639 ± 7	0.70263	1.757	0.6380	0.2196	0.513430 ± 8	0.51339	15.4 ± 2	--	--
MSI79C ¹	3.45	1.59	0.11953	0.0574	7.174	0.0231	0.703162 ± 8	0.70315	0.2721	0.0973	0.2163	0.513599 ± 8	0.51355	18.7 ± 2	--	--
WTL-1 ²	2.64	0.528	0.12390	0.0204	4.977	0.0118	0.702564 ± 8	0.70256	0.5154	0.4348	0.5101	0.512969 ± 6	0.51291	6.5 ± 1	0.16	-0.09
WTL-2 ²	1.45	22.2	0.11703	0.0250	26.43	0.0027	0.703186 ± 6	0.70318	0.5977	0.3621	0.3664	0.513792 ± 4	0.51375	22.5 ± 1	--	--
WTL-3 ²	2.02	0.473	0.12152	0.0054	0.372	0.0417	0.702766 ± 28	0.70275	0.4002	0.5571	0.8418	0.513203 ± 6	0.51310	11.0 ± 1	0.13	0.01
OU45852 ¹	2.35	0.172	0.12380	0.0048	0.221	0.0631	0.704426 ± 10	0.70441	0.0416	0.0453	0.6580	0.513785 ± 9	0.51370	22.4 ± 2	0.38	0.22
OU45852 ²				0.0038	0.876	0.0127	0.702057 ± 11	0.70205	0.3783	0.4270	0.6826	0.513839 ± 11	0.51376	23.4 ± 2	0.38	0.22
OU45852 ^{2#}				0.0053	1.034	0.0147	0.702061 ± 8	0.70206	0.4019	0.4392	0.6607	0.513788 ± 74	0.51371	22.4 ± 14	0.38	0.22
WRR-1 ²	1.44	4.93	0.11662	0.0037	4.853	0.0022	0.703555 ± 5	0.70356	5.101	1.445	0.1714	0.512850 ± 8	0.51283	4.1 ± 2	--	--
WRR-5 ²	0.76	9.93	0.11797	0.0220	43.75	0.0015	0.702786 ± 10	0.70279	2.860	0.0145	0.0031	0.512841 ± 6	0.51284	4.0 ± 1	--	--
WRR-7 ²	1.00	6.77	0.11653	0.0182	19.69	0.0027	0.703168 ± 9	0.70317	2.274	0.5986	0.1592	0.512993 ± 7	0.51298	6.9 ± 1	--	--
WRR-9 ³	1.06	21.0	0.11764	1.392	7.444	0.5407	0.702975 ± 7	0.70285	0.4856	0.0860	0.1071	0.512933 ± 17	0.51292	5.7 ± 3	--	--
DPP-1 ²		0.562*		0.0903	21.43	0.0122	0.701973 ± 8	0.70197	1.786	1.020	0.3454	0.513751 ± 14	0.51372	21.7 ± 3	1.14	0.70
DPP-1 ^{2#}				0.2013	19.84	0.0293	0.701971 ± 7	0.70197	1.723	0.9721	0.3412	0.513659 ± 34	0.51363	19.9 ± 7	1.08	0.61
DPP-2 ¹	1.34	22.8*	0.11559	0.3567	5.706	0.1808	0.703029 ± 6	0.70299	0.2490	0.0522	0.1268	0.512865 ± 8	0.51285	4.4 ± 2	--	--
DPP-3 ²	3.24	0.270*		0.0168	57.99	0.0008	0.702466 ± 10	0.70247	3.464	1.455	0.2585	0.513412 ± 5	0.51339	15.1 ± 1	1.90	0.89
DPP-3 ^{2#}				0.0235	57.50	0.0012	0.702464 ± 7	0.70246	3.404	1.479	0.2582	0.513414 ± 7	0.51339	15.1 ± 1	1.92	0.90

Sample ^A	Al ₂ O ₃ ^B (wt %)	La/ Yb _N ^C	¹⁸⁷ Os/ ¹⁸⁸ Os _(i) ^D	Rb (ppm)	Sr (ppm)	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr ^E		Sm (ppm)	Nd (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd ^E		ϵ_{Nd} ^F	T _{CHUR} (Ga) ^G	T _{DM} (Ga) ^G
							measured	initial				measured	initial			
DPP-5 ²		0.043*		0.1632	18.18	0.0260	0.702112 ± 9	0.70211	2.088	1.157	0.3351	0.513393 ± 8	0.51336	14.7 ± 2	0.83	0.31
DPP-5 ² #				0.2320	19.09	0.0351	0.702076 ± 5	0.70207	2.083	1.170	0.3397	0.513386 ± 14	0.51336	14.6 ± 3	0.80	0.29
DPP-6 ¹	3.87	0.945*	0.12566	0.4554	7.389	0.1782	0.702818 ± 7	0.70279	0.5864	0.1978	0.2040	0.512918 ± 4	0.51290	5.5 ± 7	--	--
WFP-1 ²	0.98	10.8	0.13293	0.2501	120.0	0.0060	0.702725 ± 3	0.70272	1.230	0.2151	0.1058	0.512894 ± 9	0.51288	5.0 ± 2	--	--
WFP-2 ²	1.96	0.762	0.12684	0.0204	20.45	0.0029	0.702811 ± 4	0.70281	1.750	0.6860	0.2371	0.513037 ± 5	0.51301	7.8 ± 9	1.50	-0.73
WFP-4 ²	0.56	7.09		0.0256	11.78	0.0063	0.703213 ± 10	0.70321	1.335	0.2487	0.1126	0.512797 ± 6	0.51279	3.1 ± 1	--	--
WFP-4 ² #				0.0223	13.45	0.0048	0.703173 ± 7	0.70317	1.528	0.2921	0.1156	0.512805 ± 6	0.51279	3.2 ± 1	--	--
WFP-5 ²	0.81	3.64		0.4626	15.26	0.0877	0.703046 ± 8	0.70303	1.067	0.2955	0.1675	0.512804 ± 6	0.51279	3.2 ± 1	--	--
WFP-8 ²	1.89	0.328	0.12354	0.0346	0.561	0.1783	0.704153 ± 19	0.70411	0.0418	0.0189	0.2732	0.514014 ± 46	0.513987	26.9 ± 9	2.73	2.20
WFP-8 ² #				0.0444	0.603	0.2133	0.704202 ± 17	0.70416	0.0407	0.0191	0.2836	0.514226 ± 19	0.51420	31.0 ± 3	2.77	2.33
WFP-11 ¹	0.54	14.5	0.12573	0.3884	1.524	0.7371	0.702916 ± 7	0.70292	0.1320	0.0277	0.1269	0.512879 ± 9	0.51287	4.7 ± 2	--	--
WFP-11 ²				0.3947	73.66	0.0155	0.702680 ± 7	0.70268	4.737	0.9747	0.1244	0.512856 ± 11	0.51284	4.2 ± 2	--	--
Chatham Islands																
P80180 ²	0.78	14.3	0.12408	1.8972	284.1	0.0193	0.702925 ± 9	0.70290	34.04	7.837	0.1392	0.512829 ± 4	0.51275	3.7 ± 8	--	--
P80290 ¹	0.57	20.0		0.1991	5.471	0.1052	0.703025 ± 5	0.70290	0.1342	0.0279	0.1256	0.512816 ± 6	0.51275	3.5 ± 1	--	--
P80291 ²	0.64	23.2	0.12989	0.7867	341.1	0.0067	0.702902 ± 6	0.70289	27.49	5.601	0.1232	0.512815 ± 9	0.51275	3.5 ± 2	--	--
P80354a ¹	0.61	9.01		0.0659	2.510	0.0759	0.703014 ± 5	0.70295	0.1034	0.0292	0.1706	0.512836 ± 15	0.51277	3.9 ± 3	--	--
P80354b ¹	0.51	4.59	0.12169	0.0865	1.471	0.1702	0.703187 ± 5	0.70304	0.0926	0.0280	0.1829	0.512862 ± 17	0.51279	4.4 ± 3	--	--
P80354c ¹	0.74	2.94	0.12163	0.0698	4.047	0.0499	0.702875 ± 4	0.70283	0.6535	0.195	0.1806	0.512843 ± 5	0.51277	4.0 ± 1	--	--

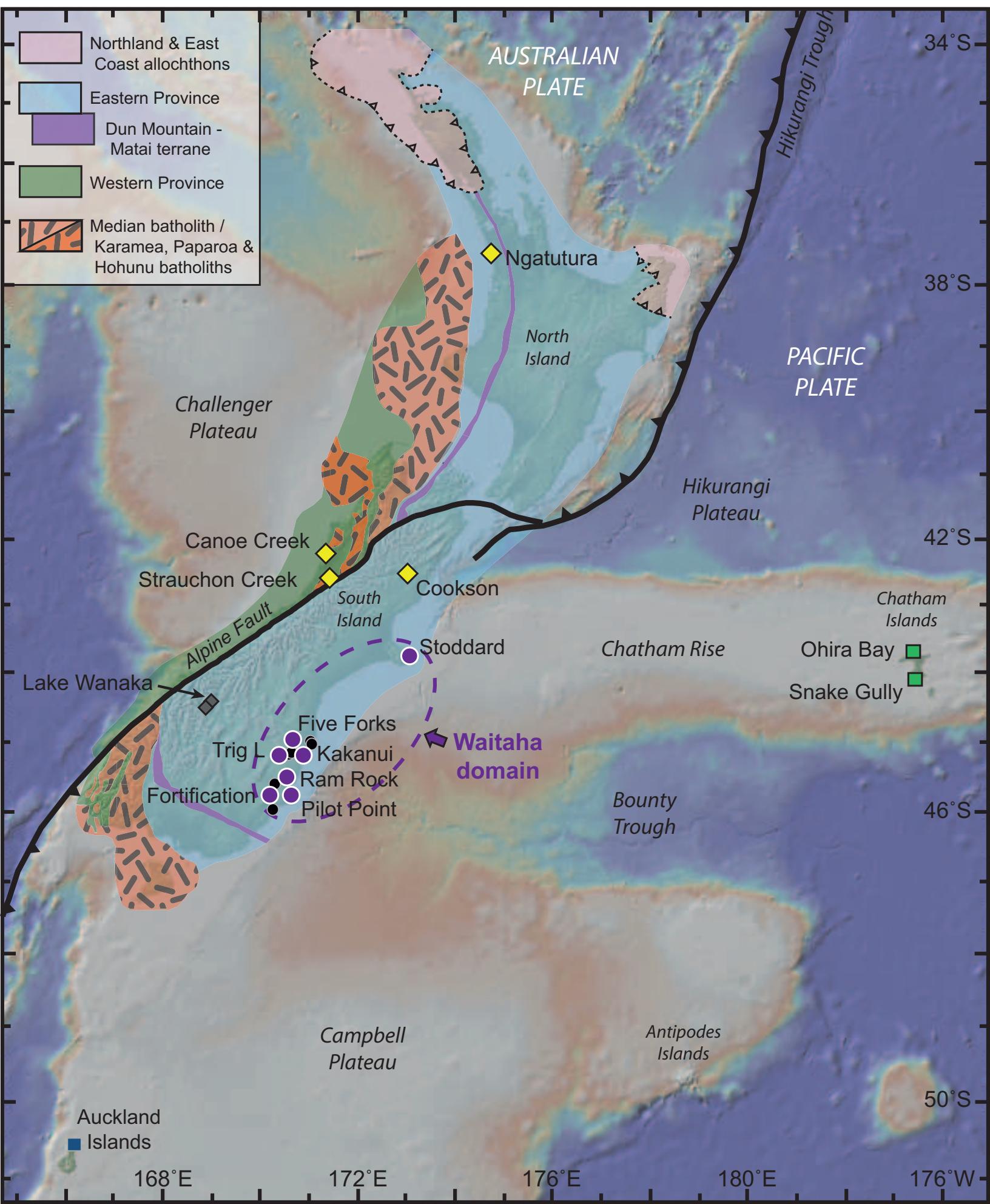
Table 2: U-Th-Pb concentration and isotopic data from New Zealand mantle xenoliths.

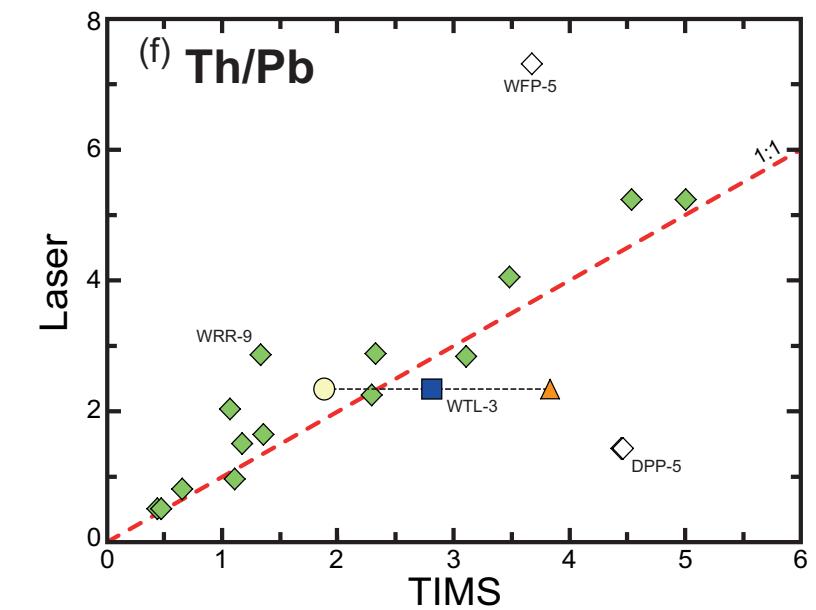
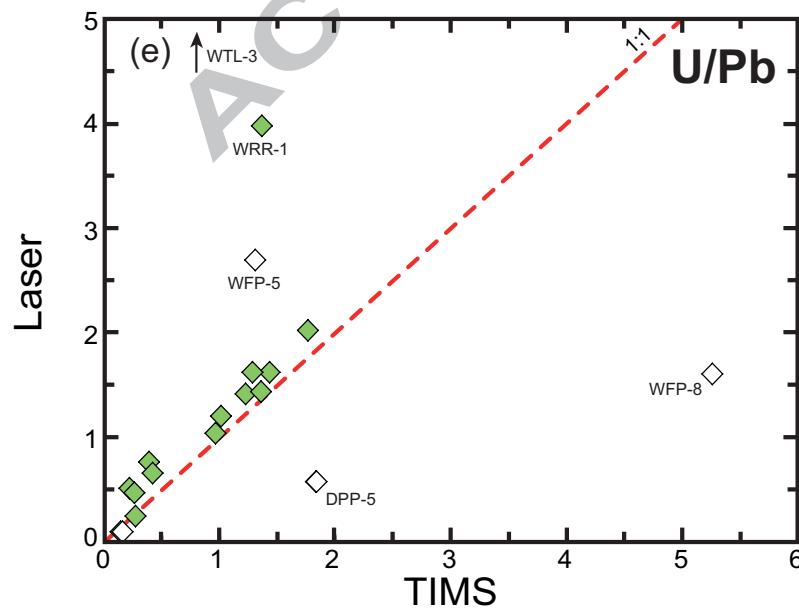
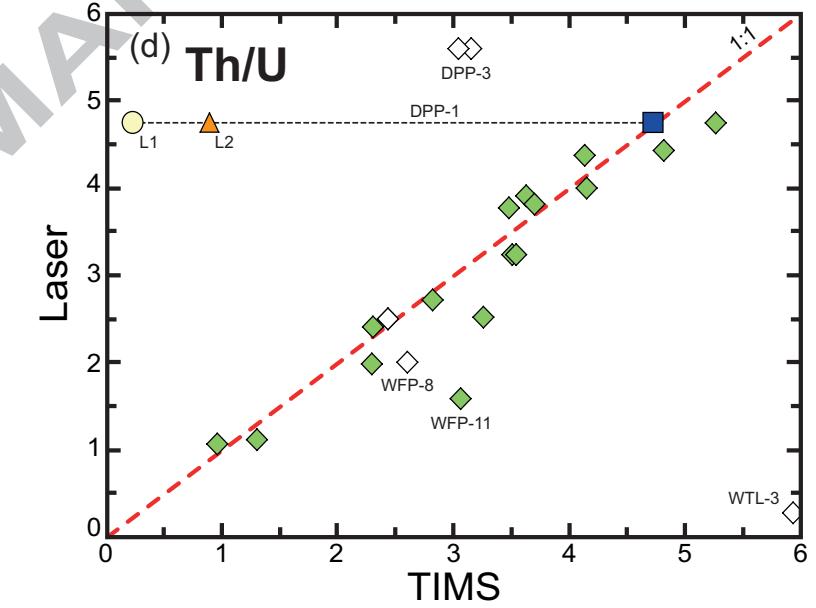
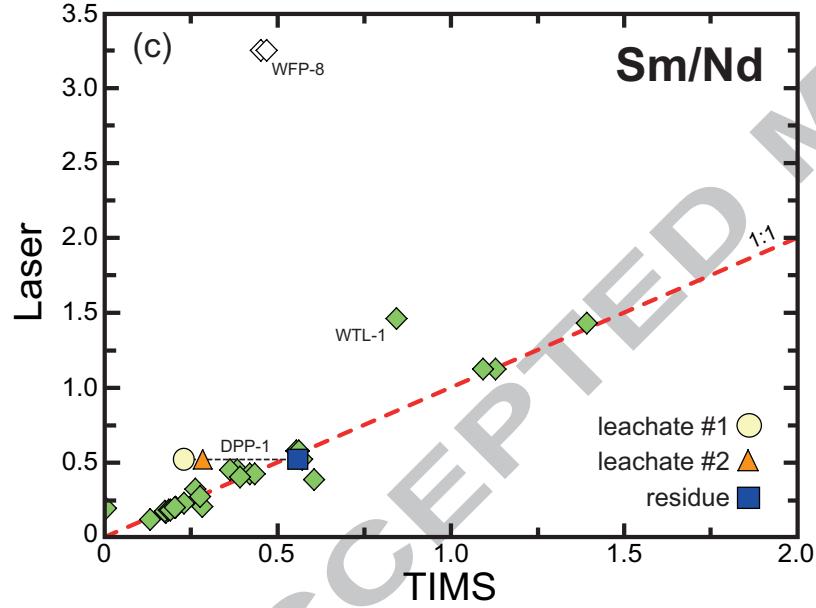
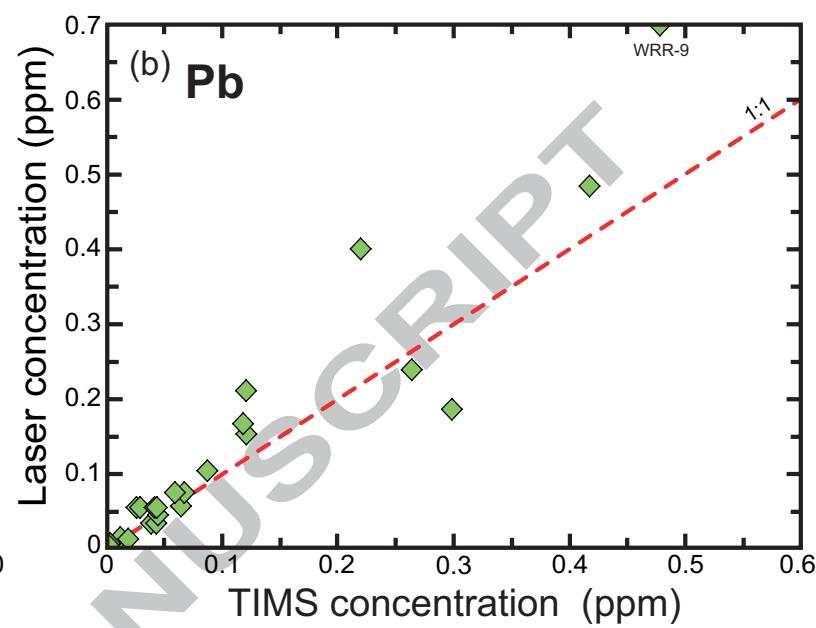
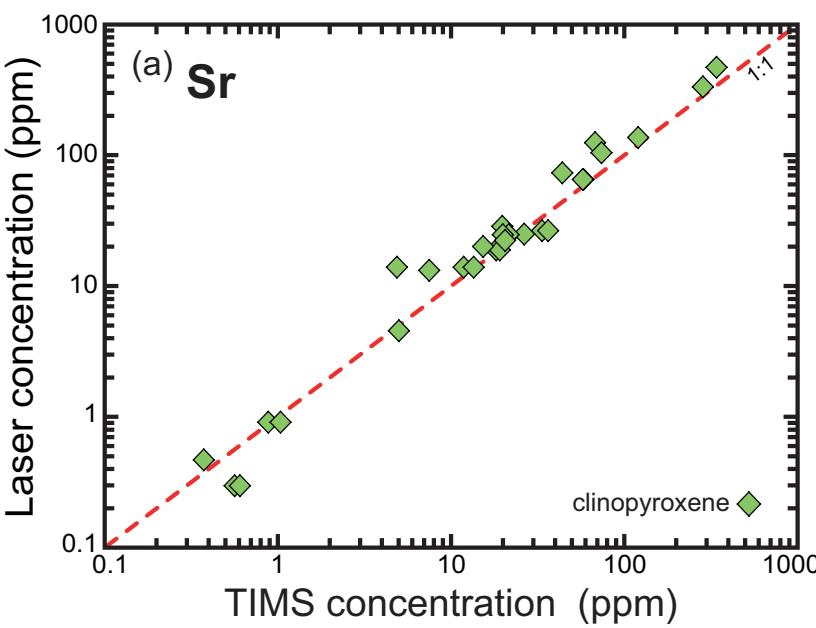
Sample ^A	U	Th	Pb	$^{238}\text{U}/^{204}\text{Pb}$	$^{235}\text{U}/^{204}\text{Pb}$	$^{232}\text{Th}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{204}\text{Pb}$ ^B		$^{207}\text{Pb}/^{204}\text{Pb}$ ^B		$^{208}\text{Pb}/^{204}\text{Pb}$ ^B		$\Delta 7/4$ ^C	$\Delta 8/4$ ^C	T_{Meta}^{206} ^D (Ma)	T_{Meta}^{208} ^D (Ma)
	(ppm)	(ppm)	(ppm)				measured	initial	measured	initial	measured	initial				
North Island																
NGB3 ¹	0.0009	0.0017	0.0284	1.99	0.014	3.95	18.681 ± 6	18.68	15.5853 ± 57	15.59	38.461 ± 15	38.46	6.93	25.0		
NGB4 ¹	0.0289	0.1308	0.0290	63.2	0.458	291.1	18.638 ± 2	18.62	15.5575 ± 22	15.56	38.2950 ± 60	38.27	4.62	13.5		
South Island																
P43153b ¹	0.0111	0.0155	0.0223	32.0	0.232	45.4	19.55 ± 10	19.15	15.463 ± 96	15.44	38.33 ± 31	38.15	-14.7	-93.1		
P45280 ¹	0.0004	0.0013	0.0144	1.92	0.014	6.01	19.000 ± 7	18.97	15.6358 ± 61	15.64	38.81 ± 16	38.79	8.52	21.8		
MSI33D ¹	0.0105	0.0352	0.0295	23.6	0.171	80.1	20.063 ± 5	19.97	15.6694 ± 35	15.66	39.6731 ± 94	39.57	0.35	-21.0		
MSIK33C ¹	0.0062	0.0154	0.0083	48.8	0.354	123.1	19.839 ± 11	19.64	15.7051 ± 68	15.70	39.217 ± 21	39.06	6.36	-39.5		
<i>Waitaha domain</i>																
MSI20C ²	0.4589	1.6653	0.4785	63.1	0.458	232.9	19.909 ± 2	19.83	15.6247 ± 9	15.621	39.6110 ± 37	39.51	-2.44	-8.55	34.7	48.1
MSI20G ¹	0.0186	0.0660	0.0166	73.5	0.533	266.0	19.937 ± 10	19.84	15.6381 ± 68	15.634	39.603 ± 19	39.49	-1.41	-12.8	32.3	41.5
FvF-1 ^{2c}	0.0544	0.1910	0.0382	93.7	0.680	334.7	20.015 ± 8	19.75	15.6205 ± 44	15.608	39.435 ± 13	39.14	-4.02	-39.1	30.7	22.8
FvF-1 ^{2f}	0.0548	0.1940	0.0428	84.1	0.610	303.3	20.006 ± 7	19.77	15.6242 ± 50	15.613	39.433 ± 14	39.16	-3.55	-38.1	33.5	25.1
MSI79C ¹	0.0095	0.0334	0.0132	47.5	0.345	169.9	20.150 ± 13	19.90	15.6523 ± 85	15.641	39.457 ± 23	39.17	-2.30	-53.1	78.5	47.6
WTL-1 ²	0.0005	0.0003	0.0441	0.69	0.005	0.52	19.687 ± 4	19.69	15.6156 ± 23	15.616	39.1155 ± 70	39.12	-0.95	-31.4	--	--
WTL-2 ²	0.1129	0.1470	0.0642	115.5	0.838	153.0	20.121 ± 6	19.79	15.6405 ± 27	15.625	39.3013 ± 91	39.16	-3.16	-65.2	30.8	32.3
WTL-3 ²	0.0004	0.0026	0.0009	31.6	0.229	190.9	21.13 ± 23	21.04	15.6334 ± 725	15.63	39.40 ± 24	39.23	-14.8	-176.9	311	36.7
OU45852 ¹	0.0003	0.0019	0.0065	2.58	0.019	19.1	19.679 ± 8	19.67	15.6144 ± 52	15.61	39.165 ± 19	39.15	-0.99	-25.4	--	--
OU45852 ^{1#}	0.0002	0.0004	0.0026	5.33	0.039	10.1	19.760 ± 26	19.75	15.5962 ± 197	15.60	39.208 ± 52	39.20	-3.68	-30.9	--	--
OU45852 ²	0.0004	0.0014	0.0116	2.44	0.018	8.24	19.568 ± 10	19.56	15.6089 ± 48	15.61	39.056 ± 17	39.05	-0.33	-22.8	--	--
WRR-1 ²	0.1640	0.6070	0.1205	90.4	0.656	340.6	20.252 ± 2	20.02	15.6701 ± 10	15.66	39.9903 ± 34	39.71	-1.62	-12.1	48.6	55.3
WRR-5 ²	0.3213	0.3081	0.2638	80.8	0.586	78.8	20.384 ± 5	20.18	15.6666 ± 44	15.66	39.688 ± 14	39.62	-3.40	-58.3	64.8	161
WRR-7 ²	0.1633	0.3748	0.1207	89.4	0.648	208.6	20.176 ± 4	19.95	15.6518 ± 24	15.64	39.5720 ± 70	39.40	-2.63	-44.8	43.8	49.9
WRR-9 ³	0.1140	0.3969	0.2986	25.2	0.183	89.2	20.122 ± 2	20.06	15.6583 ± 18	15.66	39.5315 ± 58	39.46	-1.39	-42.3	140	107
DPP-1 ²	0.2104	0.9960	0.0257	552.2	4.00	2659	20.699 ± 47	19.59	15.5809 ± 335	15.53	40.648 ± 91	38.94	-15.4	-0.47	13.2	12.1
DPP-1 ^{2#}	0.1990	1.0485	0.0288	466.2	3.38	2499	20.693 ± 9	19.75	15.6396 ± 60	15.60	40.635 ± 17	39.03	-9.45	-0.95	15.5	12.8
DPP-2 ¹	0.0129	0.0224	0.0115	74.4	0.540	130.8	20.068 ± 13	19.92	15.6463 ± 70	15.64	39.510 ± 23	39.43	-2.00	-37.9	43.2	70.0
DPP-3 ²	0.0057	0.0180	0.0414	8.39	0.061	26.9	17.260 ± 8	17.24	15.3692 ± 56	15.37	36.867 ± 15	36.85	0.72	37.6	--	--

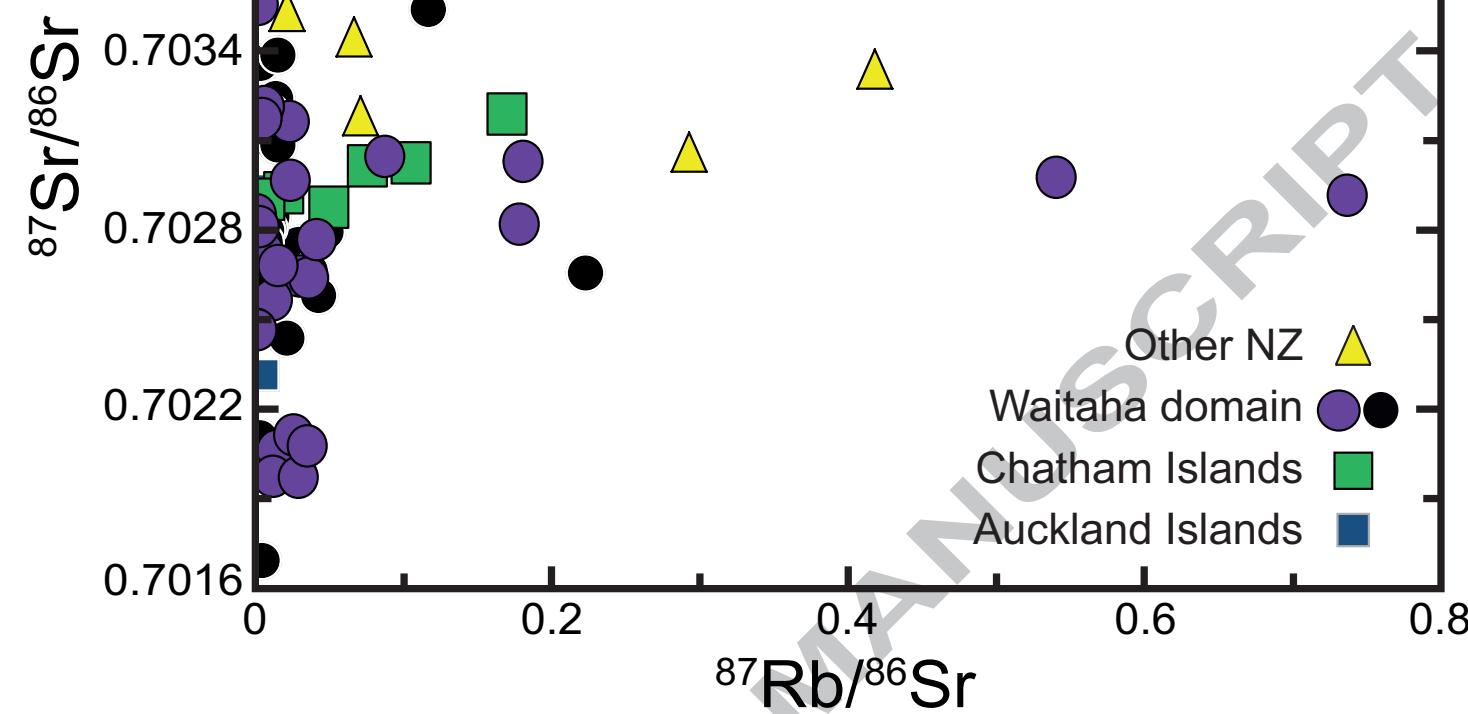
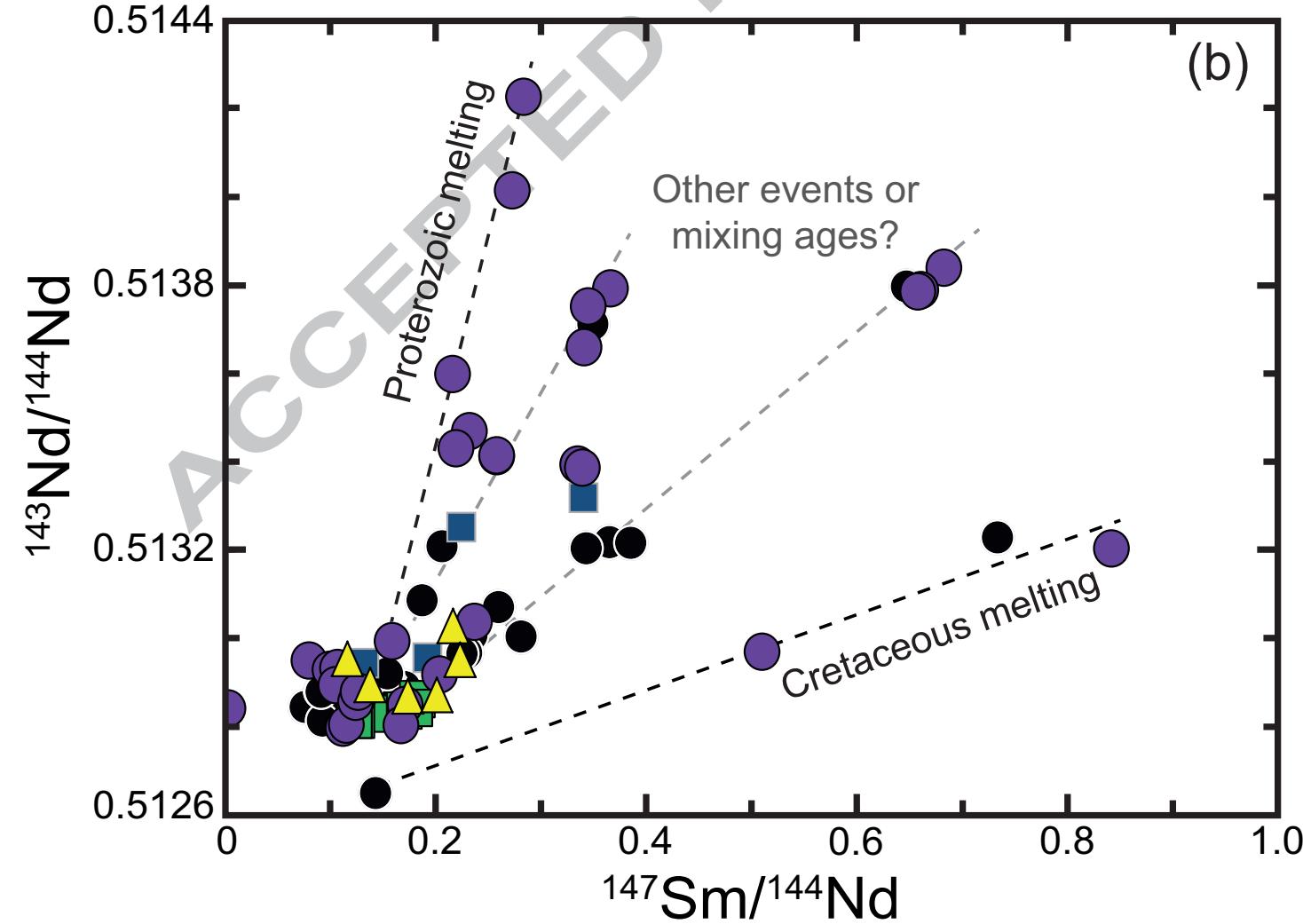
Sample ^A	U (ppm)	Th (ppm)	Pb (ppm)	$^{238}\text{U}/^{204}\text{Pb}$	$^{235}\text{U}/^{204}\text{Pb}$	$^{232}\text{Th}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{204}\text{Pb}$ ^B		$^{207}\text{Pb}/^{204}\text{Pb}$ ^B		$^{208}\text{Pb}/^{204}\text{Pb}$ ^B		$\Delta 7/4$ ^C	$\Delta 8/4$ ^C	T_{Meta}^{206} (Ma) ^D	T_{Meta}^{208} (Ma) ^D
DPP-3 ² #	0.0066	0.0202	0.0431	9.44	0.068	29.2	17.469 \pm 5	17.45	15.3662 \pm 40	15.37	37.078 \pm 11	37.06	-1.84	33.1	--	--
DPP-5 ²	0.0049	0.0119	0.0027	113.4	0.822	280.7	18.00 \pm 24	17.77	15.36 \pm 20	15.36	37.42 \pm 49	37.24			--	--
DPP-5 ² #	0.0054	0.0131	0.0029	114.8	0.833	284.3	18.443 \pm 25	18.21	15.358 \pm 20	15.35	37.664 \pm 52	37.48	-13.2	-26.1	--	--
DPP-6 ¹	0.0187	0.0594	0.0237	52.5	0.380	169.6	20.380 \pm 7	20.27	15.6794 \pm 41	15.67	39.816 \pm 12	39.71	-2.08	-45.1	99.0	90.3
WFP-1 ²	0.4092	1.9719	0.0670	472.9	3.43	2318	25.420 \pm 17	24.28	15.9347 \pm 39	15.88	47.278 \pm 26	45.50	-31.2	91.8	79.3	71.5
WFP-2 ²	0.0156	0.0650	0.0588	17.6	0.127	74.2	20.157 \pm 7	20.11	15.6769 \pm 39	15.68	39.618 \pm 13	39.56	0.09	-37.8	212	152
WFP-4 ²	0.1187	0.2735	0.1178	67.2	0.487	157.6	20.598 \pm 4	20.43	15.6815 \pm 22	15.67	39.9391 \pm 69	39.82	-4.19	-58.6	97.7	113
WFP-5 ²	0.0242	0.0684	0.0186	88.1	0.639	252.8	21.008 \pm 21	20.80	15.6948 \pm 66	15.69	40.588 \pm 30	40.39	-7.35	-43.8	105	122
WFP-8 ²	0.0079	0.0205	0.0015	356.7	2.59	944.7	21.34 \pm 10	20.48	15.684 \pm 39	15.65	40.42 \pm 14	39.70	-11.9	-100.2	31.9	29.1
WFP-11 ¹	0.0081	0.0250	0.0217	24.6	0.178	77.6	20.436 \pm 7	20.38	15.6939 \pm 47	15.69	39.763 \pm 14	39.70	-1.24	-57.2	224	183
WFP-11 ²	0.0184	0.0565	0.0869	14.2	0.103	44.1	20.585 \pm 28	20.55	15.7134 \pm 91	15.71	39.975 \pm 39	39.94	-0.90	-53.9	447	416
Chatham Islands																
P80180 ²	0.0911	0.2972	0.2197	27.9	0.202	92.5	20.801 \pm 11	20.44	15.7590 \pm 34	15.74	40.235 \pm 15	39.85	1.31	-54.1	279	256
P80290 ¹	0.0403	0.1176	0.0173	163.0	1.18	484.5	22.433 \pm 14	20.29	15.8369 \pm 38	15.74	41.817 \pm 18	39.80	-8.59	-93.2	112	115
P80291 ²	0.1074	0.4444	0.4175	16.7	0.121	70.5	19.514 \pm 13	19.29	15.6708 \pm 75	15.66	39.130 \pm 25	38.84	6.44	-8.93	--	21.2
P80354a ¹	0.0049	0.0092	0.0167	19.2	0.139	36.7	19.914 \pm 6	19.72	15.6863 \pm 27	15.68	39.2346 \pm 92	39.12	3.66	-46.8	115	97.9
P80354b ¹	0.0037	0.0094	0.0103	23.6	0.171	60.9	20.041 \pm 11	19.81	15.7217 \pm 59	15.71	39.435 \pm 19	39.25	5.83	-42.1	128	125
P80354c ¹	0.0050	0.0108	0.0222	14.8	0.108	32.4	19.919 \pm 7	19.77	15.7126 \pm 45	15.71	39.182 \pm 13	39.08	6.24	-52.7	151	78.1

Table 3: Errorchron age estimates of Zealandia lithospheric metasomatism

Locality	Type ^A	238U-206Pb system		235U-207Pb system		232Th-208Pb system		n ^B	Weighted Average	Basis for exclusion ^C
		Age (Ma)	MSWD	Age (Ma)	MSWD	Age (Ma)	MSWD			
Chatham Islands	All-inclusive:	110 ± 54	1810	121 ± 81	22	115 ± 52	172	6 (6)	114 ± 33	Small scale heterogeneity
	Preferred:	110 ± 9.3	20	119 ± 77	7.5	115.9 ± 3.1	2.8	4	115.3 ± 2.9	
Fortification Peak	All-inclusive:	54 ± 36	415	51 ± 51	28	63 ± 24	184	7 (6)	59 ± 18	Preserved ancient heterogeneity
	Preferred:	72 ± 10	350	74 ± 23	15	67.1 ± 5.7	54	6	68.6 ± 4.7	
Pilot Point	All-inclusive:	31 ± 33	9820	40 ± 83	360	22 ± 19	1590	8 (5)	25 ± 16	Lack of equilibrium: whole rock vs. cpx
	Preferred:	41.5 ± 9.1	220	68 ± 32	30	28.3 ± 3.5	33	6	30 ± 18	
Trig L	All-inclusive:	30 ± 110	86	36 ± 17	0.91	25 ± 21	3.9	6 (4)	32 ± 13	Large offset in $^{206}\text{Pb}/^{204}\text{Pb}$
	Preferred:	26 ± 15	71	36 ± 17	1.16	24 ± 31	5.1	5	30 ± 10	

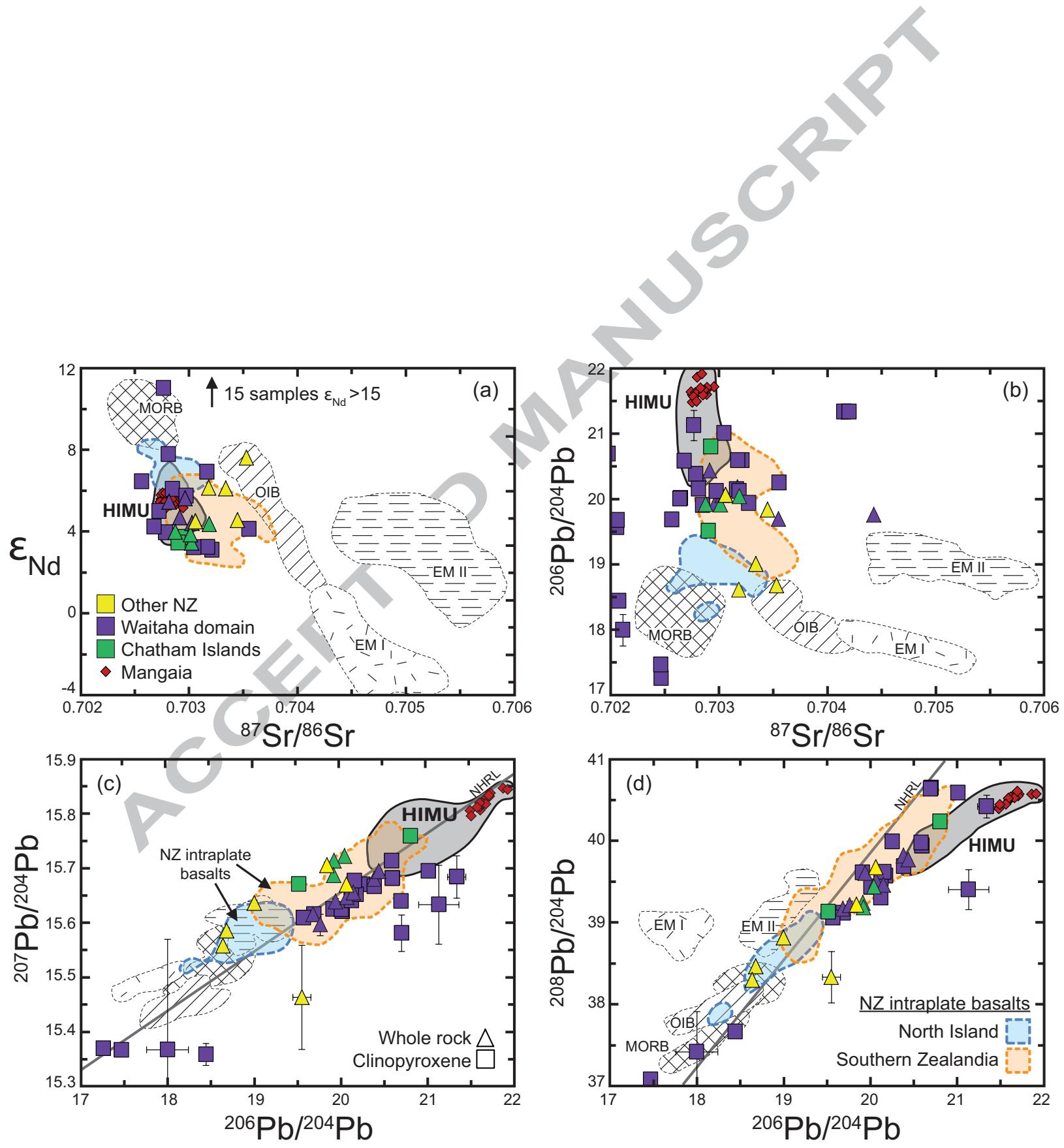


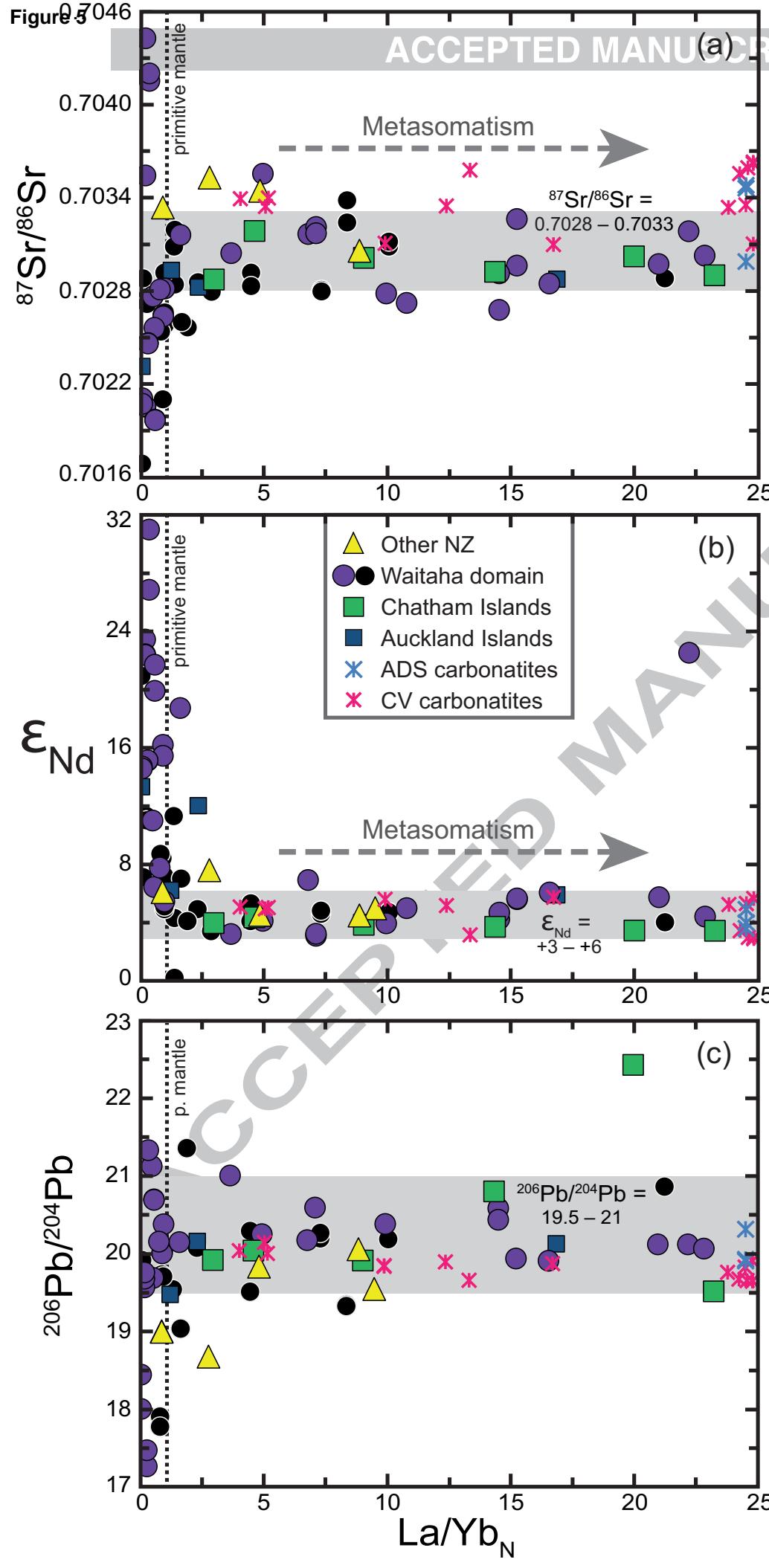


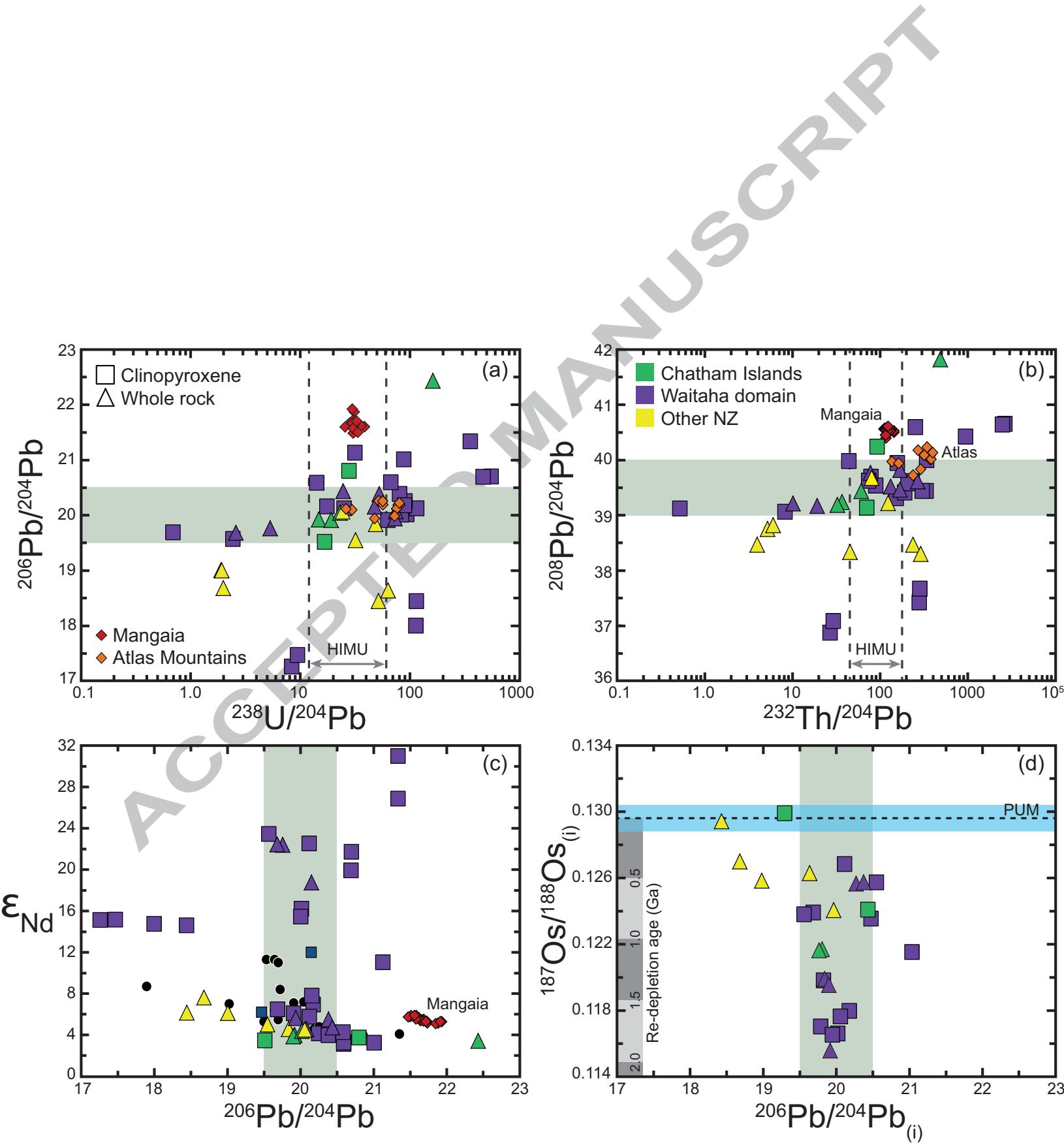
 $^{87}\text{Rb}/^{86}\text{Sr}$ Other events or
mixing ages?

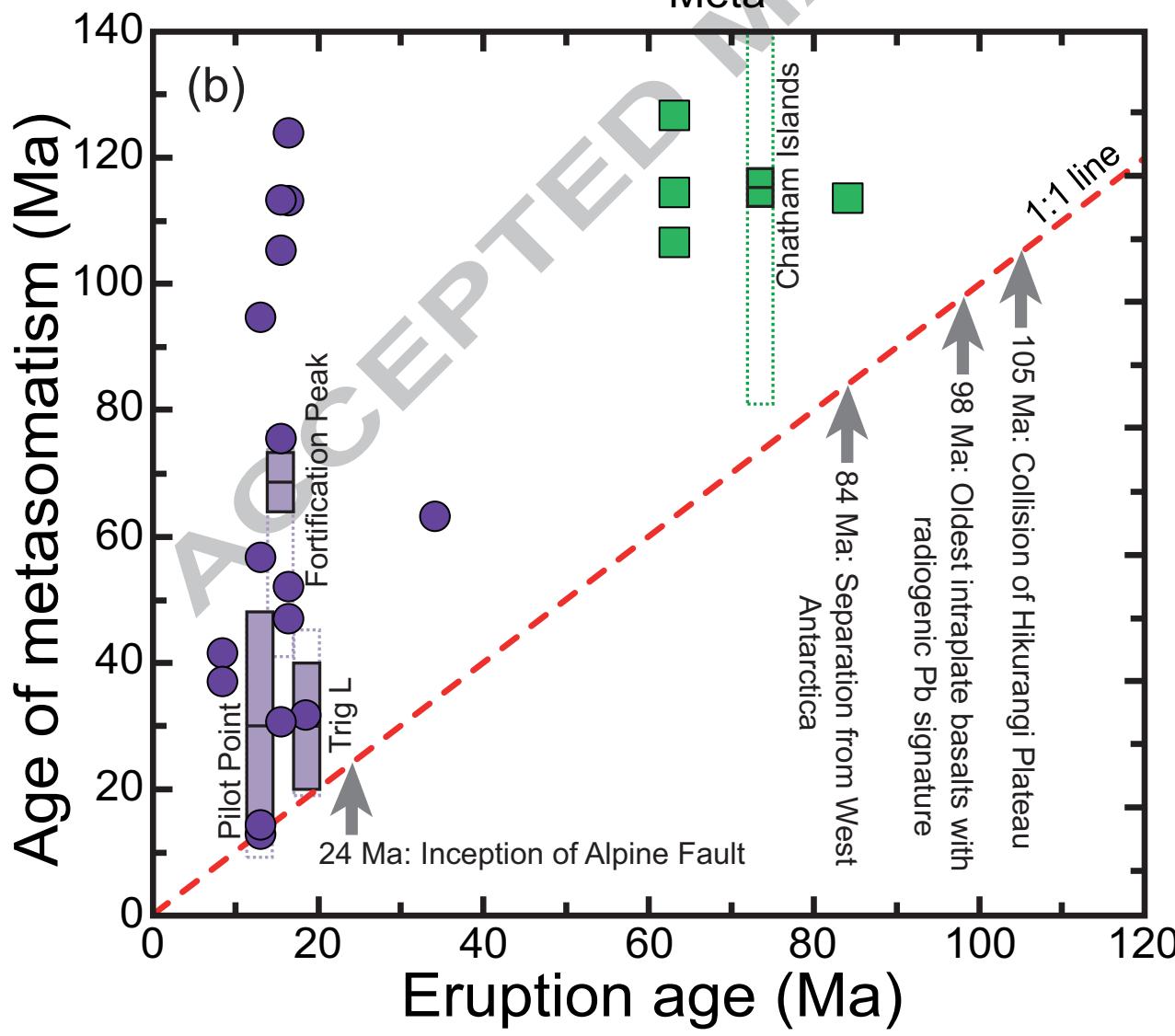
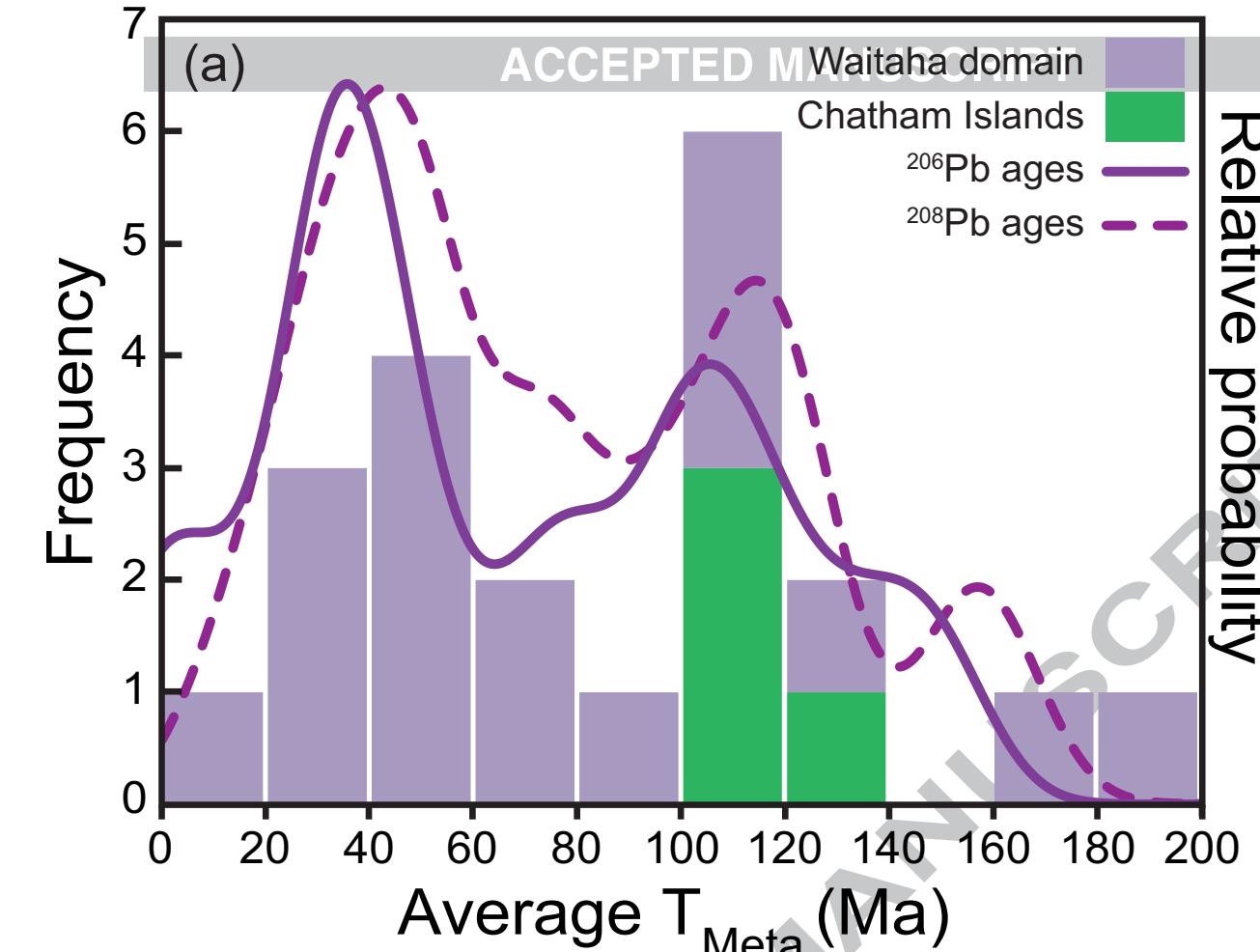
Proterozoic melting

Cretaceous melting









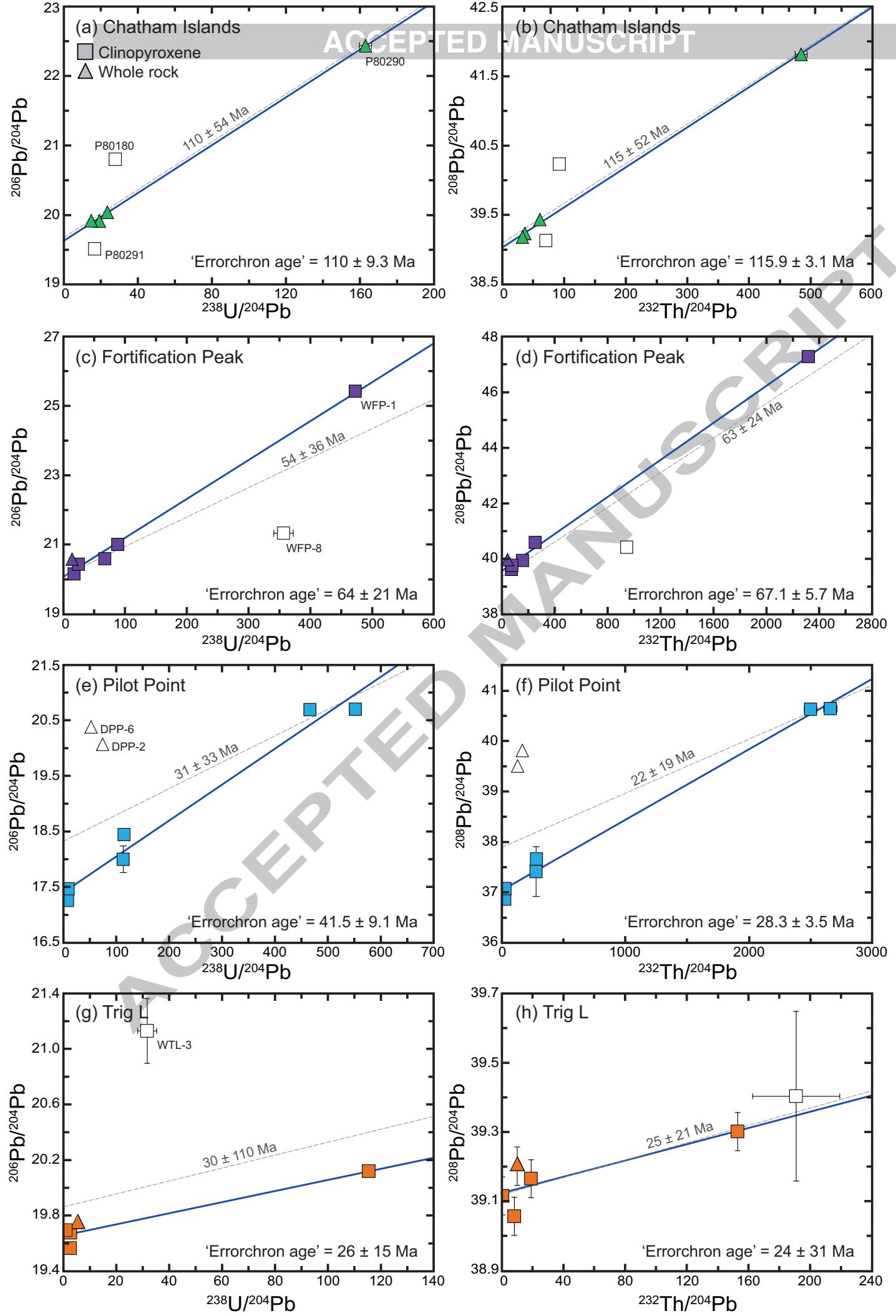
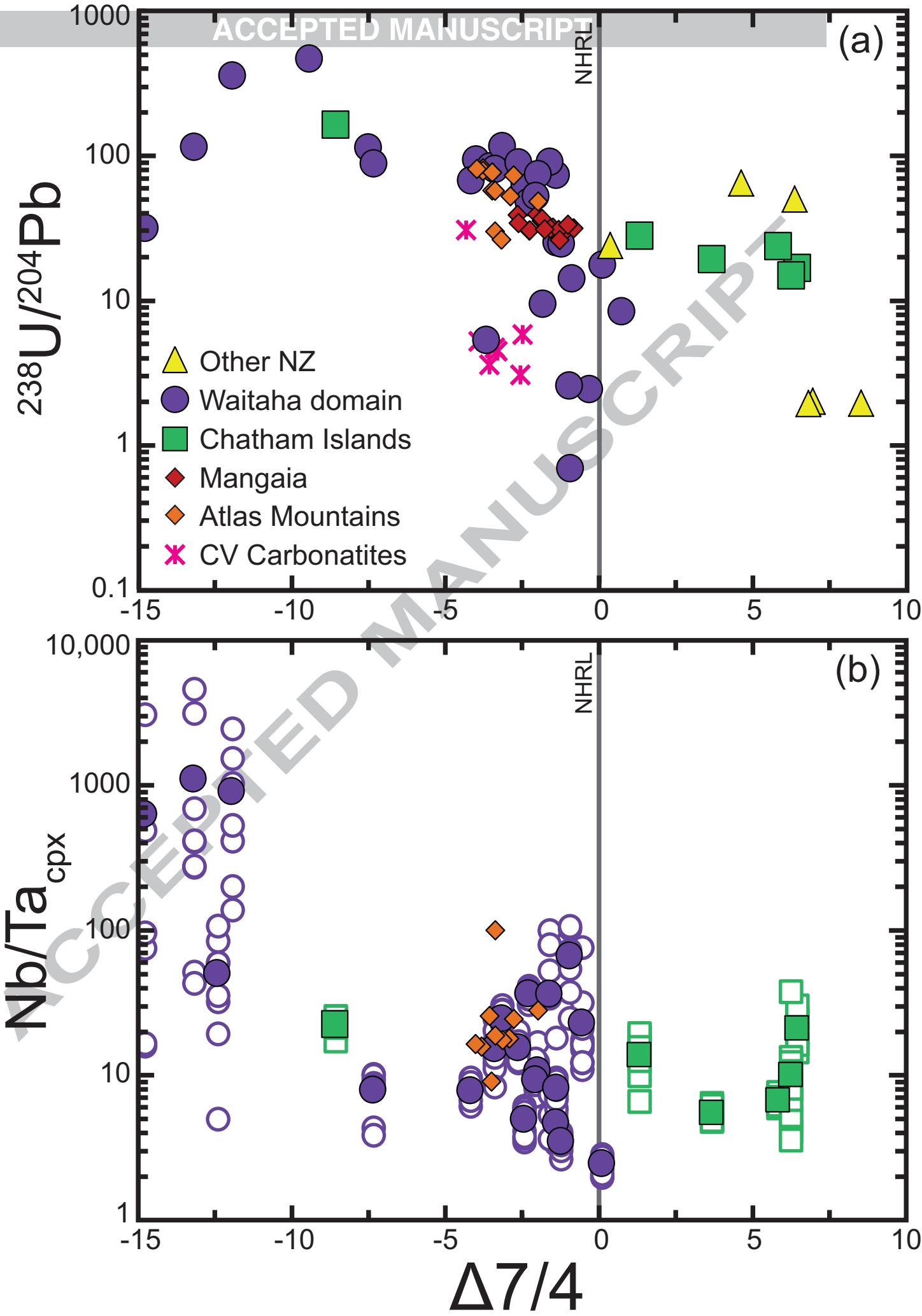
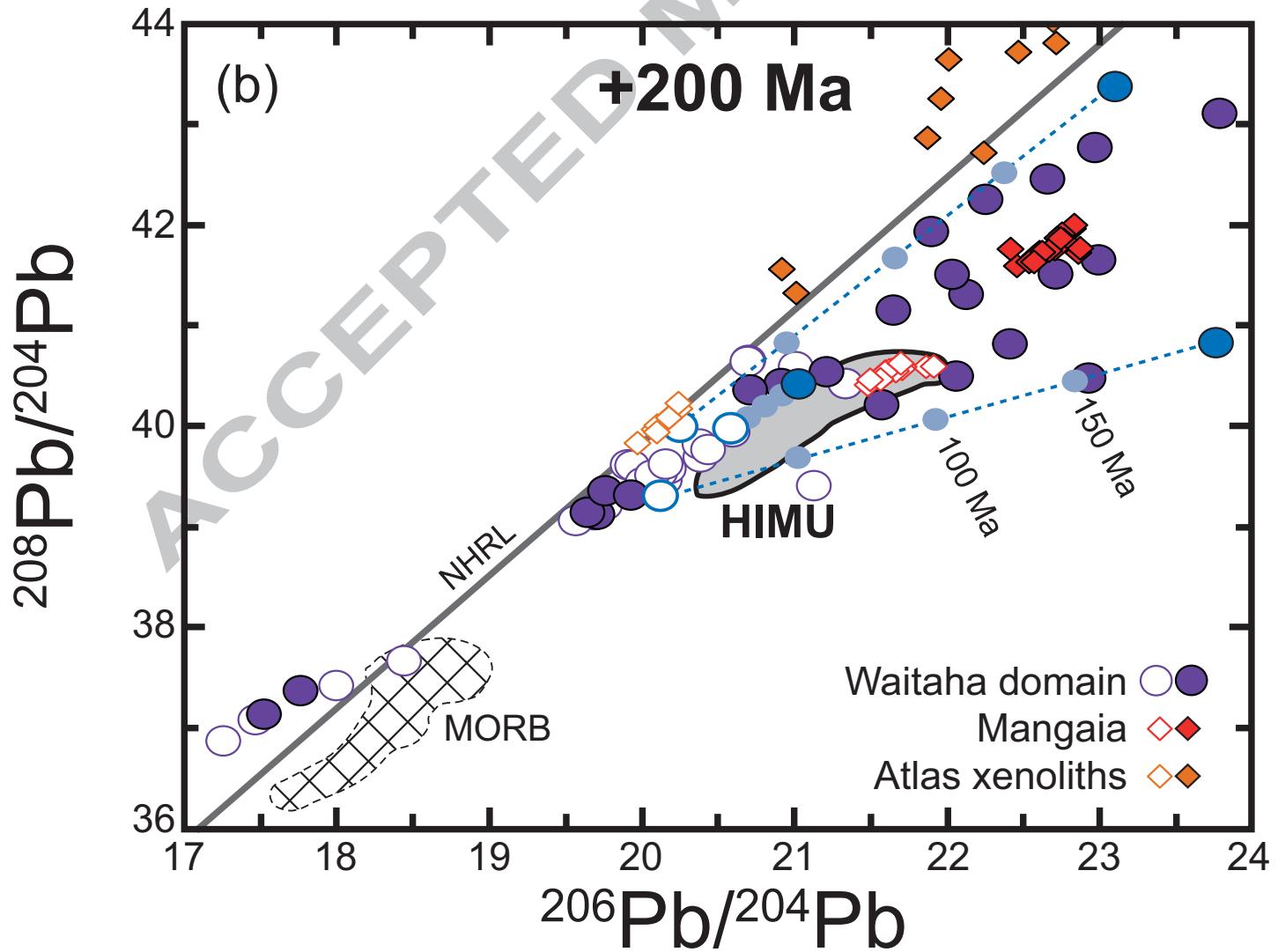
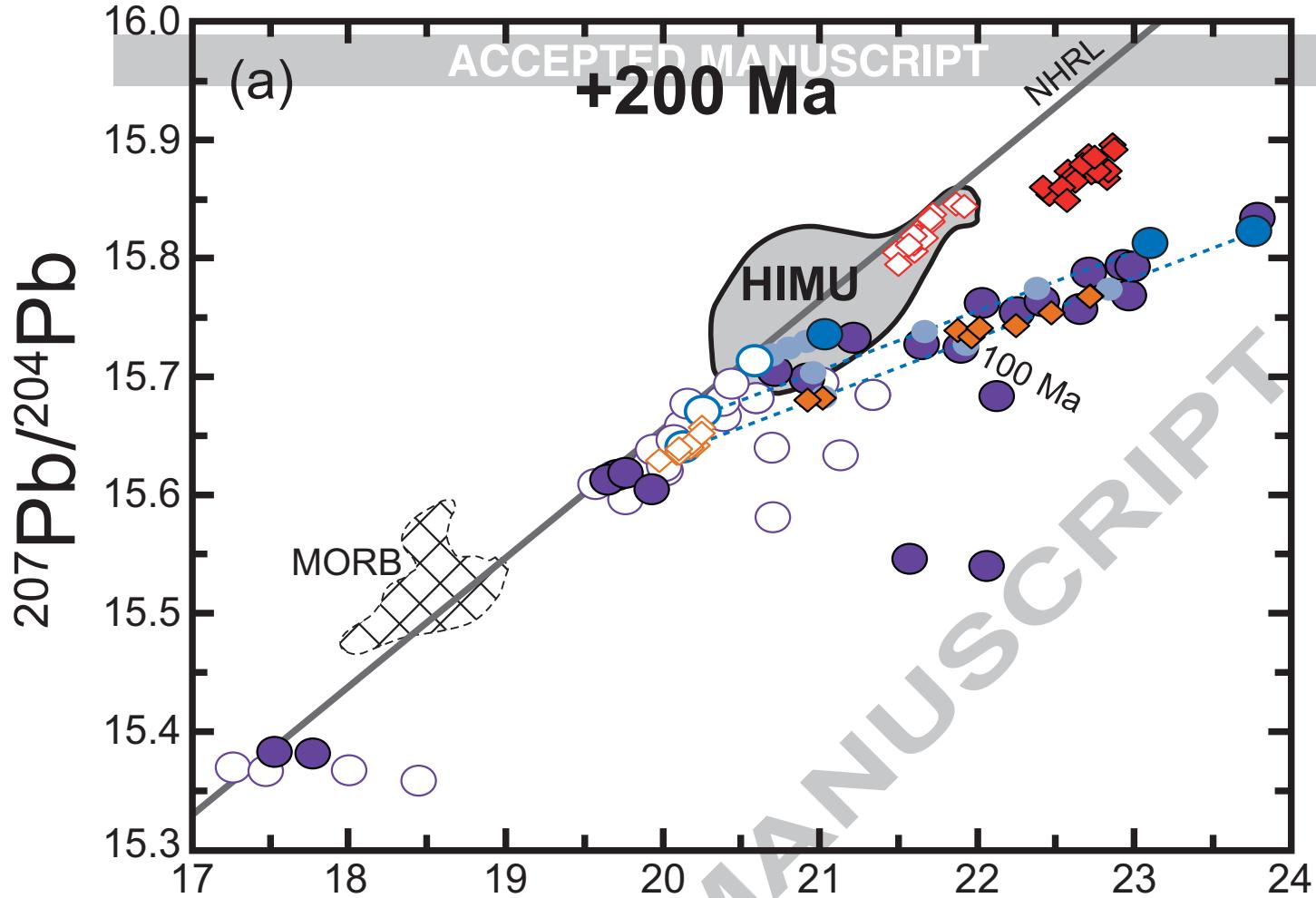
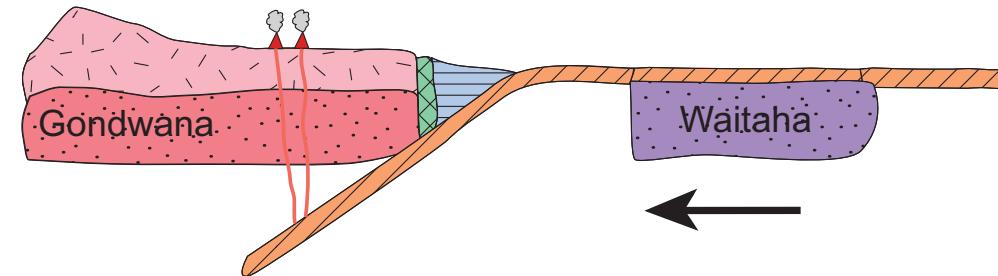


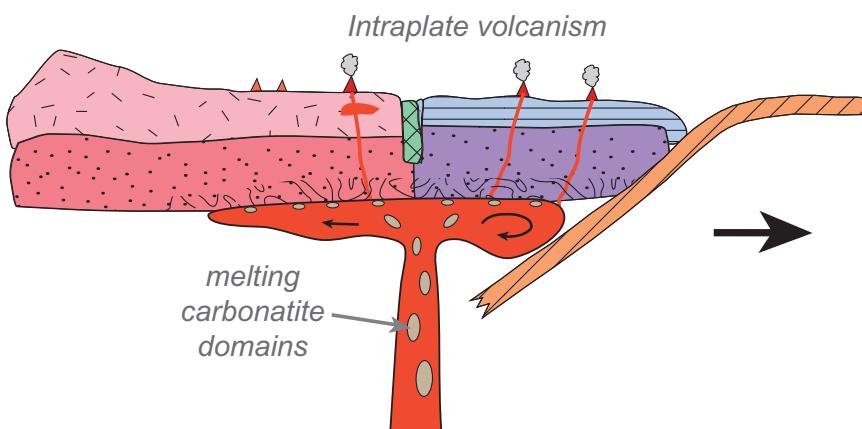
Figure 9



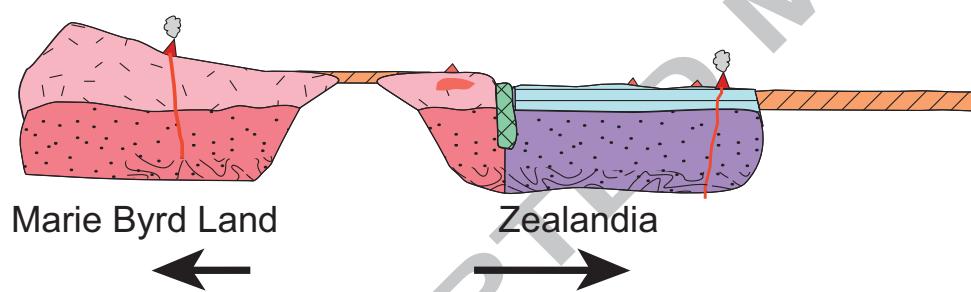




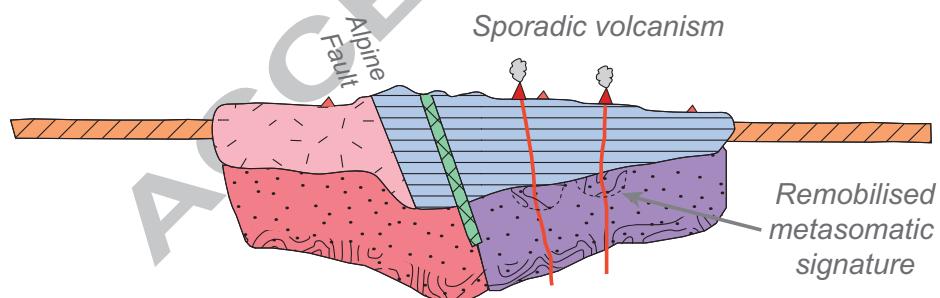
(b) Metasomatism of the lithospheric mantle at ca. 120-110 Ma



(c) Separation of Zealandia from Marie Byrd Land at ca. 84 Ma



(d) Ongoing intraplate volcanism and metasomatism of the SCLM



Oceanic crust



Continental crust



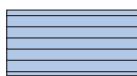
Lithospheric mantle



Waitaha domain



Dun Mountain ophiolite belt



Eastern Province sedimentary terranes



Metasomatised SCLM