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1 **The lithospheric mantle and lower crust/mantle relationships under Scotland: a**
2 **xenolithic perspective**

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17

18 **Abstract:** In the British Isles the majority of volcanic rocks containing upper mantle and
19 lower crustal xenoliths occur in Scotland. Most of the occurrences are of Carboniferous-Permian age.
20 This paper presents new data on the mineral chemistry of spinel lherzolite xenoliths from the five
21 principal Scottish tectonic terranes. Compositional variations among the four principal minerals
22 emphasize the broad lateral heterogeneity of the sub-continental lithospheric mantle across the region.
23 The remarkable range of Al₂O₃ vs CaO exhibited by the clinopyroxenes demonstrates the extremely
24 complex tectono-magmatic history of the Scottish lithosphere. The generalised age increase from
25 southern and central Scotland to the older Northern Highland and Hebridean Terranes of the north and
26 north-west, with concomitant complexity of geological history is reflected also by trace element and
27 isotopic studies. Reaction relationships in lherzolites from the Hebridean Terrane, due to pervasive
28 metasomatism, involve secondary growth of sodic feldspar. This, and light REE-enrichment of
29 clinopyroxenes points to involvement of a natro-carbonatitic melt.

30 Most pyroxenitic xenoliths are inferred to have formed as cumulates from underplating
31 magmas, forming a basal crustal layer with a generally sharp discontinuity above the underlying
32 (dominantly lherzolithic) mantle. A second discontinuity is inferred to separate these ultramafic
33 cumulates from overlying, broadly cognate meta-gabbroic cumulates.

34

35 Xenoliths of upper mantle and lower crustal origin occur over a wide area of
36 the British Isles but are most abundantly represented in Scotland. There, some
37 seventy xenolith localities are distributed in a crudely arcuate pattern that embraces a)
38 southern and central Scotland (the Palaeozoic Southern Uplands and Midland Valley
39 Terranes), b) the western sides of the Proterozoic Grampian Terrane (between the
40 Highland Boundary Fault and the Great Glen Fault), c) the western and north-eastern
41 margins of the Northern Highland Terrane (bounded by the Great Glen Fault and the
42 Moine Thrust), d) the Archaean-Palaeoproterozoic Hebridean Terrane west of the
43 Moine Thrust. Thus material from beneath all five of the major tectonic terranes can
44 be sampled (Fig. 1). Almost all of the hosts (basanites and other sub-silicic mafic
45 rocks) are of late Palaeozoic age (lower Carboniferous to Permian) with the exception
46 of a dyke near Loch Roag, Outer Hebrides (Hebridean Terrane) dated at c. 47 Ma
47 (Menzies *et al.*, 1988; M. Timmerman, pers. com.).

48

49

Fig. 1 about here.

50

51 The generalised age increase of the shallow crust from SE to NW across
52 Scotland probably reflects that of the deeper crust and lithospheric mantle (Menzies &
53 Halliday, 1988; Macdonald & Fettes, 2007). Furthermore, the composition of the
54 sub-continental lithospheric mantle (SCLM) may itself be broadly related to the
55 tectonic-thermal age of the overlying crust (Griffin *et al.*, 2009).

56

57

58

Previous accounts of the Scottish SCLM have largely concerned their
geochemistry, focussing on incompatible elements and isotopic characteristics
(Menzies *et al.*, 1987; Menzies *et al.* 1988; Menzies & Halliday, 1988; Long *et al.*

59 1991; Downes *et al.*, 2001; 2007; Bonadiman *et al.*, 2008.) These accounts have
60 drawn attention to the extensive lateral chemical heterogeneity in the SCLM as
61 represented by mantle xenoliths. The last petrological overview of the Scottish SCLM
62 was compiled over twenty years ago (Hunter & Upton, 1987). The present paper
63 considers how this heterogeneity is reflected in the mineral chemistry of the spinel
64 lherzolite xenoliths from thirteen localities distributed across the five tectonic
65 terranes. It presents new data on the spinel lherzolites, considers the implications of
66 previous geochemical studies and discusses the probable relationships between the
67 mantle and crust.

68 Late Palaeozoic extensional tectonics affecting Scotland led to widespread
69 intraplate mafic volcanism typified by the OIB-type geochemistry common to such
70 (typically) alkalic basaltic provinces (Smedley, 1988; Wallis, 1989; Upton *et al.*,
71 2004; Kirstein *et al.*, 2006). Carboniferous volcanism in central Ireland, northern
72 England and southern Scotland represents one of the earliest, if not *the* earliest,
73 examples of this type of province, as exemplified by the Basin and Range, Pannonian
74 Basin and the Massif Central, that was to become globally widespread throughout
75 Mesozoic-Cenozoic times. These occurrences also provide very early instances of
76 xenoliths hosted by alkali basaltic (basanites grading to olivine melanephelinitic
77 types) rather than by kimberlite, carbonatite, lamproite or ultramafic lamprophyre
78 magmas. This non-uniformitarian phenomenon has implications for the thermal
79 history and the evolution of the continental lithosphere at large.

80 With regard to with the metamorphic zonation imposed during the peak of the
81 Caledonian Orogeny (~460 Ma), the late Palaeozoic xenolith localities in Scotland
82 tend to lie around the low-temperature isotherms, with basanitic magmatism (and
83 associated xenoliths) being absent within the higher-temperature metamorphic zones.

84 Lithospheric composition, temperatures, thickness and pre-existing structures may
85 have an important control in this geographic distribution. Woodcock and Strachan
86 (2000) noted that geophysical studies of the upper mantle beneath the British Isles
87 suggest that in some areas the main mantle fabrics may lie parallel to Caledonian and
88 Variscan structures in the overlying crust.

89 The modal distribution of xenolith types at individual localities is also
90 instructive. Some occurrences yield a wide spectrum of xenoliths, representing both
91 mantle and lower crustal rock types but at others the compositional range is more
92 restricted. Thus at some, peridotite appears to be the sole rock-type whilst at others
93 only pyroxenitic \pm meta-gabbroic xenoliths occur.

94 The xenoliths are typically small (<30 mm) so that composite xenoliths are
95 rare. Apart from scarce quartzo-feldspathic xenoliths of (assumed) mid-crustal
96 derivation (<20 km depth), most fall into one of the following categories: a)
97 peridotitic, with metamorphic fabrics, b) pyroxenitic (*sensu lato*), including wehrlites,
98 olivine clinopyroxenites, clinopyroxenites, websterites grading to orthopyroxenites,
99 and metasomatised varieties such as biotite- and pargasite/kaersutite pyroxenites
100 grading to glimmerites or hornblendites and c) meta-gabbroic to meta-dioritic
101 xenoliths, typically possessing granulite-facies fabrics ('mafic granulites'). Most of
102 the pyroxenitic (s.l.) and mafic xenoliths are inferred to have originated as cumulates
103 from basaltic magmas (e.g. Upton *et al.*, 1983; Halliday *et al.*, 1993; Downes *et al.*
104 2007). Data from several well-studied Scottish xenolith suites point to a generalised
105 genetic relationship between the pyroxenites and mafic granulites. Although some of
106 the pyroxenites may have crystallised within the upper mantle, the majority are
107 considered to be representative samples of the deepest crustal layers (Downes *et al.*,

108 2001; Downes *et al.*, 2007; Upton *et al.*, 1998 and 2001). This conclusion is supported
109 by fluid inclusion studies (Kirstein *et al.*, 2004).

110 Despite the inference that many of the pyroxenitic xenoliths originated as
111 cumulates, some may have been generated by metasomatism of peridotitic protoliths
112 (*cf.* Meen, 1987) as has been argued by Menzies and Halliday (1988). The very scarce
113 garnet pyroxenite xenoliths are mainly confined to the Midland Valley (Upton, *et al.*,
114 2003) and are regarded as high-pressure (~20 kbar) crystallisation products of alkali
115 basalt magmas.

116

117 **Mantle peridotites**

118 The Scottish mantle xenoliths almost exclusively comprise spinel lherzolite,
119 the commonest of all mantle-derived rocks found in xenoliths and orogenic peridotite
120 bodies and stable to depths of 55-60 km (Griffin & O'Reilly, 1987a). Although
121 traditionally considered as restitic (Dick & Bullen, 1984; Arai, 1994), the lherzolites
122 may, alternatively, represent re-fertilised products of previously depleted harzburgitic
123 rocks (Bodinier *et al.*, 2008; Griffin *et al.*, 2009).

124 Whereas true harzburgite xenoliths are unknown from Scotland,
125 clinopyroxene is scarce in the most depleted lherzolites (e.g. at Oldhamstocks in the
126 Southern Uplands terrane). The lherzolites exhibit granoblastic to porphyroclastic
127 textures, with rare flaser textures. Olivines are typically 10-100 micron diameter
128 whilst orthopyroxene porphyroclasts can attain 10mm. The Scottish lherzolite
129 xenoliths tend to be finer-grained in comparison with others on a global basis (B.E.
130 Tabor, pers.com.), possibly due to cataclastic deformation during Caledonide
131 tectonism with a lack of subsequent annealing by the late Palaeozoic.

132 Rare late-stage phlogopite or pargasite in the lherzolites denotes overt or
133 modal metasomatism (Alexander *et al.*, 1986; Hunter & Upton, 1987). Whilst
134 chrome-diopside and spinel typically appear unaltered, olivines and orthopyroxenes
135 have commonly experienced alteration to serpentine minerals or carbonate and at only
136 a few localities are the peridotite xenoliths substantially pristine.

137 Representative analyses for olivines, orthopyroxenes, clinopyroxenes and
138 spinels are presented in Tables 1 & 2. Analyses of feldspar are given in Table 3.
139 Electron microprobe analyses were carried out at four laboratories. In Edinburgh, the
140 instrument used was a Cameca SX100, equipped with 5 x wavelength dispersive
141 (WD) spectrometers. It was operated at a gun potential of 20 kV and a probe current
142 (as measured with a Faraday Cup) of 20 nA. Counting times on the elements were 20
143 seconds on both peak and background. Elemental calibration was made against a set
144 of well-established standard materials. The instrument used at the Istituto di
145 Geoscienze e Georisorse in Italy was a Cameca Camebax equipped with 4 x WD
146 spectrometers. The gun potential used was 15 kV with a current of 20 nA. 100
147 second count times were used for analysis. At the Vrije Universiteit, Amsterdam, a
148 JEOL JXA-8800M electron probe was used with a 15 kV accelerating voltage and 25
149 nA current. In Vienna the equipment was a Cameca SX100 with 4 WDS and an EDS.
150 An accelerating voltage of 15 kV with a 20 nA focused beam current and a 20 second
151 counting time on peak positions were used for routine analyses. For feldspars, a 5
152 micron defocused beam current and a counting time of 10 seconds for Na and K were
153 used. For trace element analysis in olivine (Al, Cr, Ti, Zn, P) the applied conditions
154 were: accelerating voltage 20 kV, beam current 150 nA and counting times of 150 and
155 100 seconds on peak and background positions respectively. Detection limits were:
156 Al, 15ppm; Ti, 29 ppm; Ca, 26 ppm; Ni, 79 ppm; Mn, 55 ppm, Cr, 33 ppm; Zn, 55

157 ppm; P, 43 ppm). Matrix corrections on all instruments were made by the proprietary
158 PAP software supplied by the manufacturer of the instruments

159

160 *Table 1 about here*

161 **Olivines**

162 The olivines range from Fo₉₁ to Fo₈₇ with the majority clustering around Fo₈₉
163 (Table 1), as is typical for spinel lherzolites globally (*cf.* Basaltic Volcanism Study
164 Project, 1981). Spinel lherzolites have, until recently (Bodinier *et al.*, 2008; Griffin *et*
165 *al.*, 2009) been regarded as restites that have undergone variable depletion as a
166 consequence of Proterozoic or early Palaeozoic melt extraction events. A MgO vs.
167 NiO wt% plot for the olivines is shown in Fig.2. The data are unfortunately limited
168 because in most samples the olivines have been pseudomorphed. The diagram shows
169 MgO contents from 48 to 51 wt% and NiO from 0.33 to 0.45 wt% for the majority of
170 the analyses, mostly within the normal ranges for mantle olivines (McDonough &
171 Rudnick, 1998). Some, however, particularly from Rinibar have NiO contents <0.3
172 wt% (Table 1), possibly related to metasomatism as reported by Bonadiman *et al.*
173 (2008).

174

175 *Fig. 2 about here (MgO vs. NiO, olivines)*

176

177 CaO contents of the olivines are generally <0.07 wt% with higher values (>0.1
178 wt%) from Streap Com'laidh (Northern Highland Terrane), Hexpath & Oldhamstocks
179 (Southern Uplands Terrane) and especially in some of the Ruddon's Point and Fidra
180 (Midland Valley Terrane) xenoliths (0.11 – 0.20 wt%). Olivine Cr₂O₃ contents are

181 typically 0.01-0.02 wt% but higher values (up to 0.05 wt%) are found in the Cooms
182 Fell and Oldhamstocks lherzolites (both from the Southern Uplands Terrane).

183 The sub-Hebridean lithospheric mantle was shown to be highly enriched both
184 isotopically and in light rare-earth elements (LREE) (Menzies *et al.*, 1987; Menzies &
185 Halliday, 1988; Long *et al.*, 1991). Accordingly a special attention was given to
186 analysis of trace components in the olivines from Loch Roag. Al, however, is close to
187 the detection limit (25 ppm), Ti and P lie below the detection limit (25 ppm), with Zn
188 of ~60 ppm. The Cr level is < 0.01 wt. %. Thus none of these values suggests any
189 enrichment.

190

191 *Fig 3 about here.*

192

193 **Spinel**

194 Spinel compositions vary widely: although most have Cr# values (Cr/(Al +
195 Cr)) <20, these can be up to ~70 (Table 1). Fig. 3 depicts the relationship between Cr#
196 (spinel) and Mg# (olivine) in relation to the ol-sp mantle array (Arai, 1987; 1994).
197 Most data fall into the lower Cr# (<0.30) - lower Mg# field, suggesting that the
198 peridotites are relatively undepleted. However, spinels from Riska (Northern
199 Highlands) and some from Cooms Fell (Southern Uplands) have higher Cr# denoting
200 greater degrees of melt depletion. Whilst the Rinibar (Orkney, Northern Highland
201 Terrane) and Loch Roag (Hebridean Terrane) compositions lie both within and
202 outside the field boundaries, all have the comparatively high Cr# values indicative of
203 greater degrees of depletion (Arai, 1994). Spinel data (Cr# vs. Mg#) for the lherzolite
204 spinels (Fig. 4) support the conclusion that many xenoliths, including those from
205 Colonsay (Grampian Terrane) and Streap Com'laidh (Northern Highland) samples,

206 have experienced only relatively small degrees of melt extraction in contrast to those
207 from Machrihanish (Grampian Terrane) and many of those from Rinibar (Northern
208 Highland). The forsterite content can be significantly changed as a result of
209 metasomatism by incompatible-element-rich fluids and can be shifted off the ol-sp
210 mantle array towards lower Fo values (Irving, 1980; Goto & Arai, 1987).
211 Accordingly metasomatism may be the explanation for those Rinibar and Loch Roag
212 samples lying to the right of the array (Fig. 4).

213 The Loch Roag spinels are distinctive compositionally in having lower MgO
214 contents (15-13 wt% as opposed to 21-17 wt% in the other spinels studied) and in
215 having higher TiO₂ (~0.33 wt% vs. <0.18 elsewhere). Their Cr₂O₃ (49-35 wt%) are
216 also relatively high, the majority of the other lherzolite spinels containing <20 wt%
217 Cr₂O₃. There is notable within sample variability as shown in sample LR81, with
218 spinels having Cr # 42.12 and 67.13 (Table 1b). Texturally the Loag Roag spinels are
219 unique among the Scottish xenoliths studied in that they have undergone a reaction in
220 which sodic feldspar was produced.

221

222 *Fig. 4 about here. (Cr# vs. Mg#, spinels)*

223

224 **Clinopyroxenes**

225 Clinopyroxene is, along with spinel, the phase least affected by secondary
226 alteration. The relationship between CaO and Al₂O₃ (Fig. 5) also provides an
227 indication of depletion, the more aluminous (>5 wt% Al₂O₃) clinopyroxenes being the
228 least depleted. The clinopyroxenes show a broad 'tail' towards lower Al₂O₃, including
229 those from Oldhamstocks, Ruddon's Point, Machrihanish, Riska and many from
230 Rinibar, that may reflect increasingly refractory character. An increase in Cr₂O₃ and

231 NiO also monitors the increasing degree of refractoriness (McDonough & Rudnick,
232 1998). NiO is typically less than 0.06 wt. % but rises to ~0.09 wt% in the Cooms Fell
233 lherzolites. Generally Cr₂O₃ in the clinopyroxenes ranges between 0.4-0.7 wt% but
234 notably higher values (>1.00 wt%) are found in Riska, Streap Com'laidh and Cooms
235 Fell lherzolites but with the highest values (up to 1.7 wt%) at Oldhamstocks.

236 Clinopyroxenes from the Loch Roag samples are noteworthy in being more
237 siliceous (~ 54 wt% SiO₂) than elsewhere in Scotland (51-52 wt%), with significantly
238 higher sodium contents (2.56 – 2.32 wt% Na₂O compared to 2.0 – 0.8 wt% at other
239 localities) (Table 2). They tend also to be comparatively poor in Al₂O₃ (4.7-4.2 wt%
240 Al₂O₃) whereas most of the other lherzolite clinopyroxenes contain 8-5 wt%.

241

242 *Fig. 5 about here (Ca vs. Al₂O₃ in cpxs)*

243

244 **Orthopyroxenes**

245 TiO₂ values of the orthopyroxenes are mostly <0.12 wt% but are particularly
246 low (below detection) at Oldhamstocks (Table 2). Al₂O₃ typically falls between 2.0 –
247 4.6 wt%, rising to ~5.38 wt% at Cooms Fell. The Loch Roag orthopyroxenes are
248 relatively Al-poor (2.6 - 2.2 Al₂O₃ wt%) in comparison with values >3 wt% for the
249 majority of orthopyroxenes from other localities. CaO values generally fall into the
250 range 0.3 – 0.8 wt% but the Oldhamstocks orthopyroxenes are notably more calcic
251 (~0.98 CaO wt%) as well as being very poor in TiO₂ (Table 2).

252 Experimental and theoretical studies concerning peridotite melting provide
253 depletion trends in the form of modal abundances and mineral compositions. With
254 increasing degrees of depletion, the modal proportion of olivine increases whilst those
255 of orthopyroxene, clinopyroxene and spinel decrease. Since Al, Ca, Na, K and Ti are

256 more incompatible during melting than Mg, Cr and Ni, their concentrations are
257 highest in the least depleted peridotites. On the basis of the mineral proportions of the
258 depleted-MORB mantle (DMM) proposed by various authors (e.g. Johnson *et al.*,
259 1990, Baker & Beckett, 1999, Workman & Hart, 2005 and references therein)
260 resulting from progressive melt extraction from primitive upper mantle (PUM;
261 McDonough & Sun, 1995), we calculate the major element compositions and
262 proportions of the primitive source, assuming pressures and temperatures within the
263 spinel stability field. Using this simple model we model the melting curve for two
264 crucial elements (Al and Mg) in both ortho- and clinopyroxenes. The orthopyroxenes
265 of the Scottish peridotites broadly correspond to the hypothetical melting trend (Fig.
266 6), with the Riska and Oldhamstocks lherzolites, from the Northern Highland and
267 Southern Upland Terranes respectively, representing the most refractory mantle
268 domains. The Fidra orthopyroxenes show anomalously low values of MgO (30.5 –
269 32.5 wt%).

270

271 *Fig.6 about here (MgO vs. Al₂O₃ for opxs)*

272 The same melting curve calculated for the clinopyroxenes (Fig. 7) shows that, for
273 most of the xenolith population, this phase does not reflect the same degree of melting
274 as the coexisting orthopyroxenes. Therefore we have to postulate another process. In
275 terms of the major element compositions of the pyroxenes, only the Riska and
276 Oldhamstocks xenoliths appear to represent residues after extensive melt extraction.
277 In general, however, the clinopyroxenes cannot be described as resulting as residua
278 from simple melting. Clinopyroxenes from Ruddon's Point and Corrie na Ba have
279 particularly high Al₂O₃ contents, a characteristic considered to have come about from
280 metasomatism involving Al-rich melts (Bonadiman *et al.*, 2008).

281

282

Fig. 7. about here (MgO vs. Al₂O₃ for cpxs)

283

284 Feldspars

285

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287

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291 Isotope and trace element data

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Fig 8 about here (Nd vs. Sr data)

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305

The Loch Roag spinel lherzolites are unique among the samples studied in that corroded spinel rims are bounded by anorthoclase and greenish Na-rich clinopyroxene symplectite reaction rims. This mineral pair occurs along the sp-cpx, ol-ol and ol-cpx contacts (Hunter & Upton, 1987). The feldspar has 9.64-9.38 wt% Na₂O and 0.33-0.94 wt% K₂O (Table 3).

Apart from a few lherzolite xenoliths that have experienced modal metasomatism that generated secondary pargasite (Black Rocks (Midland Valley Terrane; Alexander *et al.*, 1986), phlogopite (Riska, Northern Highland Terrane) and sodic feldspar (Loch Roag, Hebridean Terrane), variable cryptic metasomatic enrichment has been inferred from incompatible element and isotopic data, particularly in the Northern Highland and Caledonian Foreland terranes (e.g. Menzies & Halliday, 1988). Such geochemical signatures signify marked provinciality within the lithospheric mantle a generalised decrease in enrichments from the Foreland Terrane SCLM in the NW to that beneath distinctly younger crust under the Midland Valley in the SE (Fig. 8).

Studies of the Loch Roag xenoliths have shown that the Hebridean Terrane is underlain, at least locally, by lithospheric mantle marked by extreme chemical and

306 isotopic heterogeneity (Menzies *et al.*, 1987; Menzies & Halliday, 1988; Menzies *et*
307 *al.*, 1988; Long *et al.*, 1991). The metasomatising agents were deduced to have been
308 carbonatitic melts, of probable asthenosphere origin (Long *et al.*, 1991).

309 Mantle xenoliths from Rinibar (Orkney) record two distinct metasomatic
310 styles, due to a) carbonatitic and b) more silicate-rich (kimberlitic) melts (Bonadiman
311 *et al.*, 2008). These melts were inferred to have derived from different degrees of
312 partial melting of the same source, at 550 ± 50 Ma, when both Canadian and Finnish
313 cratons were experiencing carbonatitic and kimberlitic magmatism and at a time when
314 the Rodinian supercontinent was undergoing the break-up that ultimately resulted in
315 formation of the Iapetus Ocean. Attendant on the subsequent Silurian to Devonian
316 closure of Iapetus, fluids/melts from subducting slabs are believed to have affected
317 the overlying mantle wedge and the lithospheric mantle. Such subduction-related
318 modifications are recorded in mantle xenoliths from Streap, towards the south-
319 western extremity of the Northern Highland Terrane (Menzies & Halliday, 1988;
320 Bonadiman *et al.*, 2008). The lack of a comparable subduction-related signature in
321 the Orkney xenoliths emphasizes the regional variation in metasomatic history of the
322 lithospheric mantle in northern Scotland. At all localities where radiogenic isotope
323 ratios have been measured, $^{143}\text{Nd}/^{144}\text{Nd}$ varies between 0.1524 and 0.5131 except at
324 Loch Roag in the Hebridean Terrane. $^{87}\text{Sr}/^{86}\text{Sr}$ is consistently higher in whole-rock
325 vs. clinopyroxene analyses (Fig.8).

326 The SCLM beneath the Midland Valley and Southern Upland Terranes appear
327 to have been little affected by metasomatic processes. These mantle xenoliths have
328 isotopic characteristics akin to those of OIB-source mantle (Figure 8) and closely
329 resemble the relatively depleted lithospheric mantle that underlies the Basin and
330 Range, Eifel and Massif Central (Downes, 2001). The Midland Valley experienced

331 intraplate extensional magmatism in the late Palaeozoic and the crustal thinning and
332 enhanced heat-flow may have accompanied asthenospheric upwelling, with
333 concomitant lithospheric modification (Menzies & Halliday, 1988; Downes *et al.*,
334 2001; Downes *et al.*, 2007). It may be speculated that lithospheric delamination or
335 erosion may have intervened during the period ~400-350 Ma, leaving a relatively
336 youthful Midland Valley lithosphere that was unaffected by the long-term
337 metasomatic modifications that affected the lithospheric mantle beneath the northern
338 terranes.

339 The heterogeneity of the Scottish mantle peridotites is also reflected in the
340 chondrite-normalised rare-earth element (REE) patterns of the clinopyroxenes (Fig.
341 9). None of the clinopyroxenes represent a simple melting trend and all are
342 interpreted as being “secondary”, resulting from extensive melt interaction. The light-
343 REE convexity ($La_N < Ce_N$) may have resulted from crystallisation of the
344 clinopyroxene from a primitive basaltic melt or, alternatively, from long-term-
345 interaction between the clinopyroxenes and a metasomatic melt. The latter alternative
346 appears more plausible.

347 The clinopyroxenes fall into different groups, according to the patterns of
348 REE enrichment (Fig. 9). In one group, represented by clinopyroxenes from Rinibar,
349 (Orkney) the patterns are strongly fractionated with a tendency to be concave-down
350 in their $LREE_N$, with a fall from Ce to La. The relatively low $HREE_N$ values suggest
351 that this clinopyroxene group originally equilibrated with garnet. Subsequent uplift of
352 the mantle domain may have caused progressive destabilisation of the garnet and
353 growth of spinel, redistributing the garnet REE budget between cpx and opx.
354 Considering the HREE cpx/opx distribution coefficient (*ca.* 1; GERM online data
355 bank), the ‘new’ cpx HREE profiles become less steep (although still fractionated).

356 In another group, clinopyroxenes that always have been in equilibrium with
357 spinel have almost flat or negative fractionated HREE patterns (Simon *et al.*, 2008).
358 The fertile character (high Al₂O₃ contents) and the LREE-enriched to flat chondrite-
359 normalized MREE-HREE profiles of the Streap and Fidra pyroxenes may denote
360 long-term equilibration between lithospheric mantle in the spinel stability field and
361 metasomatic melts related to subduction. Clinopyroxenes from Loch Roag lherzolites
362 are characterised by flat normalised patterns other than in the LREE where there is
363 strong enrichment from Sm to La (*cf.* Menzies & Halliday, 1988). Clearly several
364 distinct metasomatic processes have affected the Scottish lithospheric mantle in time
365 and space.

366 The meta-gabbroic (mafic granulite) xenoliths from Streap, Fidra and Loch
367 Roag, also record highly radiogenic Sr (Menzies *et al.*, 1987; Downes *et al.*, 2001)
368 and represent shallow sub-arc lithospheric mantle that was variably contaminated by
369 fluid/melts related to subduction (Upton *et al.*, 1999; Downes *et al.*, 2001, Bonadiman
370 *et al.*, 2008). This is confirmed by the range of fluid inclusion densities in the Streap
371 xenoliths with low- (0.78-0.82 g cm⁻³), high- (~ 1 g cm⁻³) density and super-dense
372 (~1.17 g cm⁻³) inclusions observed indicating fluids trapped at subcrustal depths
373 (Kirstein *et al.*, 2004).

374
375

Fig.9 about here

376

377 **Pyroxenitic xenoliths**

378 Many of the ultramafic xenoliths fall into the category of the globally
379 widespread Type 2 xenoliths (Frey & Prinz, 1978), considered to be high-pressure
380 (10–8 kb) cumulates from underplated mafic magmas (*cf.* White, 1966; Upton *et al.*
381 2001). Although in a few cases, ultramafic xenoliths interpreted as cumulates include

382 dunites, the majority of these Type 2 xenoliths contain aluminous augitic pyroxene.
383 Since pyroxene is commonly the dominant component, the xenoliths will collectively
384 be referred to as pyroxenites (*s.l.*). Most are wehrlites, olivine pyroxenites and
385 clinopyroxenites, but websterites occur and, with increasing modal orthopyroxene,
386 grade towards orthopyroxenites (Downes *et al.*, 2007).

387 Mineral data for pyroxenite assemblages are presented in Hunter & Upton
388 (1987), Upton *et al.* (1998), Upton *et al.* (2001) and Downes *et al.* (2007). The Mg#
389 values are lower than in the mantle peridotites and the olivines typically range from
390 Fo₈₇₋₇₂. In some, feldspar is a minor component, occurring either as an intercumulus
391 phase or as feldspathic schlieren, with compositions ranging from bytownite to
392 oligoclase and anorthoclase. Pargasitic to kaersutitic amphibole is common in some
393 localities, occurring as oikocrysts in igneous-textured pyroxenites or in a replacive
394 mode. Biotite is also common and gradation to glimmerite may have occurred
395 through metasomatic introduction of Fe, Ti, K and H₂O to pyroxenitic and dunitic
396 cumulates (*cf.* Lloyd, 1981). Spinels (hercynite to magnetite) are ubiquitous and
397 ilmenites, commonly geikelite-rich, also occur. Some xenoliths appear to be
398 clinopyroxene-hercynite cumulates. Although apatite is usually insignificant, it is
399 present to several modal percent in some biotite clinopyroxenites.

400 Although the textures are typically metamorphic, some xenoliths retain
401 igneous textures. These include wehrlites and olivine clinopyroxenites with
402 clinopyroxene oikocrysts enclosing early olivines, websteritic types with poikilitic
403 clinopyroxene around orthopyroxene, and hornblende pyroxenites with kaersutite
404 oikocrysts enclosing clinopyroxenes (Upton *et al.*, 2001; Downes *et al.*, 2007). In
405 these the oikocrysts pseudomorph former intercumulus melt pores between cumulus
406 olivine, orthopyroxene and clinopyroxene respectively. Some pyroxenitic xenoliths

407 have been found with intercumulus devitrification products indicating that they were
408 still above their solidus temperatures at the time of entrainment, implying a genetic
409 relationship to their host basanites.

410 Pargasite-bearing pyroxenites from Orkney, Duncansby Ness and Gribun (Mull)
411 are regarded as metasomatic products from anhydrous pyroxenites, produced by influx
412 of hydrous melts rich in Ca, Fe, Ti and K, as well as LREE, Zr, Nb, Y Ba and Sr (Upton
413 *et al.*, 1998; 2001). In extreme cases, the metasomatism culminated in production of
414 hornblendite. Differences in pyroxenitic suites from localities only a few kilometres
415 distant, (e.g. from Gribun and Fionnphort, Isle of Mull), show that such metasomatism
416 can be highly localised (Faithfull & Upton, 2006).

417 Although the majority of the pyroxenitic xenoliths appear to have originated as
418 cumulates, some may be metasomites derived from mantle peridotites. Thus, some
419 wehrlitic and olivine clinopyroxenitic xenoliths from Loch Roag were considered to
420 have resulted from metasomatic reaction between carbonatitic melts and anhydrous
421 peridotite (Long *et al.*, 1991).

422

423 **Meta-basic xenoliths**

424 Meta-gabbroic, dioritic and anorthositic xenoliths, common in all five terranes,
425 have the characteristics of the ‘mafic granulites’ widely regarded as composing the
426 continental lower crust (Rudnick, 1992; Downes, 1993). The majority are regarded as
427 having formed as plagioclase-clinopyroxene cumulates from magmas undergoing
428 fractional crystallisation and assimilation processes (Halliday *et al.*, 1993).
429 Equilibration pressure estimates of 8–6 kb for lower crustal xenoliths from Bute in the
430 western Midland Valley (Downes *et al.* 2007) agree with the other indicators of
431 relatively thin crust. Model ages, mineral ages and scattered whole-rock isochrons

432 suggest that precursors to the meta-gabbroic xenoliths were formed and
433 metamorphosed in late Proterozoic and early Palaeozoic arc terranes during the
434 Caledonian orogeny.

435 There is a remarkable overall compositional uniformity in the meta-gabbroic
436 and meta-dioritic xenoliths across all the terranes, average SiO₂, Al₂O₃ and MgO
437 contents being similar despite variation at any one locality. This uniformity implies
438 that relatively reproducible processes of assimilation and crystallisation accompanied
439 underplating by mafic magmas. Such underplating may have repeatedly exploited a
440 density plane separating ultramafic cumulates from overlying feldspathic cumulates,
441 thereby maintaining the overall structure of the lower crust (Cox, 1980; Halliday *et*
442 *al.*, 1993; Upton *et al.*, 2001).

443

444 **Crust-mantle relationships**

445 Most of western Europe is underlain by thinner, warmer and less depleted
446 lithosphere than that beneath the East European craton to which it was accreted (Gee &
447 Stephenson, 2006). New crust is thin and of relatively low density but with
448 cratonisation it gets thicker and denser and the crust/mantle boundary tends to become
449 more poorly defined (Drummond & Collins, 1986). Whereas the thickness of the
450 Scottish lithosphere is poorly constrained, it would appear probable that it thickens and
451 increases in age and complexity, as traced north-westwards from the Midland Valley
452 towards the Outer Hebrides (Menzies & Halliday, 1988).

453 Pyroxenitic dykes within the upper mantle may have provided some of the
454 pyroxenitic xenoliths (*cf.* Wilshire & Shervais, 1975; Irving, 1980; Wilshire & Pike,
455 1975; Downes, 2007). However, the observation that pyroxenites occur in abundance
456 at several Scottish localities where peridotites are rare or absent, implies that, at least in

457 those instances, they derive from sources that were not intimately associated with
458 lherzolites. We infer that the majority of the pyroxenitic xenoliths come from a (thick?)
459 ultramafic cumulate layer overlying the lherzolitic mantle and that they are themselves
460 overlain by the feldspathic (meta-gabbroic) cumulates. We regard the pyroxenitic
461 cumulates as higher-temperature magmatic products complementing the meta-gabbroic
462 cumulates much as the ultramafic basal units of large mafic intrusions do to the
463 overlying gabbroic cumulates, and that the ultramafic and mafic cumulate assemblage
464 jointly constitutes the lower crust (Upton *et al.*, 2001). Sub-horizontal boundaries
465 between the mantle lherzolites and the pyroxenitic cumulates and between the latter and
466 the meta-gabbroic cumulates are envisaged. The lower crust has long been considered
467 as consisting largely of layered igneous intrusions (e.g. Singh and McKenzie, 1993) and
468 the seismically-determined multiple layered reflectors within the mantle lithosphere
469 beneath northern Scotland (Ascencio *et al.*, 2003) are probably due to layering within
470 the cumulates.

471 The concept of a ‘petrological Moho’, marked by the discontinuity between
472 tectonised peridotitic mantle and overlying ultramafic cumulates forming the base of
473 the oceanic crust was introduced by Greenbaum (1972) and Lippard *et al.* (1986). The
474 term was used to distinguish the discontinuity between what were interpreted to be
475 restite and cumulate facies from the seismic or geophysical Moho marking the
476 boundary between ultramafic and mafic rocks. Accordingly, the seismic Moho does
477 not necessarily coincide with the base of the crustal sequence and may well lie
478 significantly above it.

479 Cox (1980) postulated that the base of the continental crust comprises
480 ultramafic cumulates overlain by gabbroic cumulates, both products of underplating
481 magmas. He suggested that a single pulse of (picritic) magma along, or close to, the

482 crust-mantle boundary would differentiate into an upper gabbroic and a lower
483 ultramafic layer, thus generating a new seismic Moho at the interface. Repeated
484 injection along the new Moho would then regenerate gabbroic cumulates above while
485 augmenting the ultramafic cumulates below. Such a discontinuity between tectonised
486 mantle peridotites and overlying pyroxenite cumulates can be considered as analogous
487 to the ‘petrological Moho’ (Lippard *et al.*, 1986; Griffin and O’Reilly, 1987b), whilst
488 that between the pyroxenitic and gabbroic zones represents the seismic Moho (Upton
489 *et al.*, 2001; Faithfull & Upton 2006). There is commonly a simple overall structure
490 in layered mafic-ultramafic intrusions in which gabbroic cumulates overlie, more or
491 less abruptly, ultramafic cumulates. Our observations are compatible with a model in
492 which a crustal sequence of ultramafic (pyroxenitic) cumulates underlies the mafic
493 cumulates.

494 There is a general acceptance that the continental seismic Mohorovičić
495 Discontinuity marks the break from intermediate or basic crustal rocks to underlying
496 ultrabasic rocks (Ringwood, 1975). Whilst the latter are frequently assumed to be
497 ‘normal’ peridotite, this ignores the xenolithic evidence that pyroxenites can play a
498 significant role in the lithospheric structure and suggests that the inferred ‘sharp’
499 transition from ultramafic to mafic cumulates may be erroneously identified as the
500 crust-mantle boundary.

501

502 **Discussion and Summary**

503 The mineral chemistry of the Scottish spinel lherzolites clearly implies
504 considerable variation in the degree to which they have undergone melt extraction
505 whereas the distinctive mineral chemistry of the Loch Roag lherzolites, together with
506 the presence of alkali feldspar reflects the metasomatism demonstrated by the LREE

507 and isotopic enrichment documented by Menzies *et al.*, (1987), Menzies & Halliday,
508 (1988) and Long *et al.*, (1991). The remarkable variation in Al₂O₃ in the
509 clinopyroxenes compared to that in individual continental xenolith provinces (Fig. 5)
510 may be attributable to the degree to which the extremely complex magmatic history of
511 Scotland (Macdonald & Fettes, 2007) has left its imprint on the lithospheric mantle.

512 The sub-Scottish lithosphere exhibits an overall increase in age (and probably
513 thickness) north-westwards beneath the Caledonides to the Archaean/Palaeo-
514 Proterozoic Terrane. Greater age provided greater opportunities for the lithospheric
515 mantle to be affected by migrating melt fractions, including those of probable
516 asthenospheric origin as well as others from subducting oceanic lithosphere. These
517 processes conferred ever greater complexity. LREE-enrichment through carbonatitic
518 metasomatism within the cratonic lithosphere of the north-west. Carbonatite-mantle
519 interaction, inferred to have taken place in the late Archaean or early Proterozoic (2–
520 2.5 Ga), enriched a depleted peridotite protolith resulting in extreme variations in Nd
521 isotopic compositions (Long *et al.*, 1991). By contrast, the relatively young
522 lithospheric mantle of the Midland Valley and Southern Uplands Terranes has much
523 greater isotopic homogeneity.

524 Mantle xenoliths within the North Atlantic and Baffin Bay Palaeogene
525 magmatic provinces appear to be restricted to very few localities and the only other
526 location known in the British Isles is Inver, Co. Donegal (Kirstein & Timmerman,
527 2000). Spinel lherzolite xenoliths from a Palaeogene dyke in west Greenland (Larsen,
528 1982) appear to provide some parallels with the Loch Roag situation in that they
529 contain alkali feldspar. In particular, they contain small amounts of phlogopite,
530 hyalophane (Na₂O 5.65 wt%: K₂O 6.54 wt%: BaO 3.13 wt%) and chlorite, the latter
531 within sieve-textured clinopyroxene, possibly pseudomorphing glass. Larsen (1982)

532 also refers to plagioclase (+ phlogopite) in another lherzolite xenolith: the plagioclase
533 (not analysed) presumably being Na-rich.

534 The presence of feldspar in the Loch Roag samples could conceivably relate to
535 a phase-change from spinel- to plagioclase-facies lherzolite, reflecting a reduction in
536 pressure (*cf.* Rampone *et al.*, 1993). Such a pressure release would be compatible with
537 the Outer Hebrides lying close to the trailing edge of the sundered continental
538 lithosphere where stretching and elevation of the SCLM would be anticipated.
539 However a) the textural relationships differ markedly from those attending a facies
540 change (Rampone *et al.*, 1993) and b) the significant content of potassium would be
541 inexplicable. Just as Larsen (1982) appealed to possible metasomatic introduction of
542 material for the genesis of the phlogopite and hyalophane in the west Greenland case,
543 so a metasomatic introduction of materials was the probable cause of the alkali
544 feldspar - green clinopyroxene association in the Loch Roag xenoliths. Our REE data
545 for the Loch Roag lherzolite clinopyroxenes amplify those of Menzies & Halliday
546 (1988) and support the deduction by Long *et al.* (1991) that the geochemical and
547 petrographic idiosyncrasies of the Loch Roag peridotites were due to percolation of a
548 carbonatitic agent. The concurrence of both alkali addition (Na>K) and LREE
549 enrichment suggests the possibility of metasomatism by natro-carbonatite melt as was
550 noted in the case of data from Grande Comore xenoliths (Coltorti *et al.*, 1999).

551 Furthermore these incompatible element enrichments (and associated isotopic
552 singularities) cannot be considered unrelated to the megacryst suite that accompanies
553 the peridotite and pyroxenite xenoliths. The megacryst suite has been shown (K-Ar
554 and Ar-Ar dating; Menzies *et al.* (1988) and M. Timmerman, pers. comm.) to be very
555 young (~47 Ma) whilst the isotopic signatures are an inheritance from ancient
556 enrichment events (Long *et al.*, 1991). The salic melts from which the megacryst

557 suite crystallised (Upton *et al*, 2009) may represent remobilised ancient and highly
558 fusible metasomatic products that ascended in advance of the mafic (host) magma in
559 the Palaeogene tectono-thermal event.

560 The pyroxenitic (*s.l.*) xenoliths are deduced to have arisen from a thick zone of
561 ultramafic cumulates overlying tectonised peridotites. The former are regarded as
562 constituting the base of the crust and the discontinuity between the two is considered
563 analogous to the ‘petrological Moho’ of ophiolitic sequences. Gabbroic/dioritic meta-
564 cumulates, broadly cogenetic to the pyroxenite cumulates, are inferred to overlie the
565 latter, separated from them by a relatively sharp discontinuity (the ‘seismic Moho’).
566 The ultramafic and mafic cumulates represent complementary products of primitive
567 basaltic magmas repetitively introduced along the density break coincident with the
568 seismic Moho during magmatic episodes.

569

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577

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803 **Figure captions**

804

805 Fig. 1. Map showing the principal tectonic terranes and xenolith localities across
806 Scotland. LR, Loch Roag; RNB, Rinibar; DCN, Duncansby Ness; STP, Streap
807 Com'Laidh; RSK, Riska; GRB, Gribun; FNP, Fionnphort; KTN, Kilchatten; ISL,
808 Islay; HN, Hawks Nib; MCR, Machrihanish; BR, Black Rocks; RP, Ruddon's Point;
809 EN, Elie Ness; FD, Fidra; Old, Oldhamstocks; HEX, Hexpath; CF, Cooms Fell.

810 Fig. 2. MgO vs. NiO wt% plotted for olivines from spinel lherzolite xenoliths from
811 Scotland.

812 Fig. 3. Cr# spinel vs. Mg# olivines from Scottish lherzolite xenoliths showing the
813 olivine-spinel mantle array (Arai, 1994).

814 Fig. 4. Cr# vs. Mg# for spinels from the Scottish lherzolite xenoliths.

815 Fig. 5. CaO vs. Al₂O₃ wt% for clinopyroxenes. **A:** data for Scottish xenolith
816 localities. **B:** comparative fields for spinel lherzolite xenoliths from other continental
817 provinces worldwide.

818 Fig. 6. MgO vs. Al₂O₃ wt% plot for orthopyroxenes from the Scottish lherzolite
819 xenoliths. PM_{opx} (large open square) is based on mass balance calculations using
820 major element composition and modal proportions (spinel stability field) for PUM
821 from McDonough & Sun (1995). The theoretical residual trend is calculated using
822 equations from Workman & Hart (2005). The trend is interrupted at the calculated
823 melting percentage for complete pyroxene consumption (opx 30%; cpx 25%).

824 Fig. 7. MgO vs. Al₂O₃ wt% for clinopyroxenes from Scottish lherzolite xenoliths. The
825 theoretical residual trend is calculated using the equations from Workman & Hart
826 (2005).

827 Fig. 8. $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ for whole-rocks and clinopyroxenes from Scottish
828 lherzolite xenoliths. Selected data from Menzies & Halliday, 1988; Downes et al.,
829 2001; 2007; Bonadiman et al., 2008.

830 Fig. 9. Chondrite-normalised REE patterns for clinopyroxene in Scottish peridotite
831 xenoliths. C1 chondrite normalization values of McDonough & Sun (1995).

Table 1b. Selected spinel analyses for key xenolith localities.

	C Fell C	CnB 51	RNB 5	Hex 1	HN 31	RP 77	Rsk 1	STP1	Fid-1	Fid-2	LR80-1	LR80-2	LR81-1	LR81-2
SiO₂	0.05	0.03	0.06	0.05	0.08	0.05	0.03	0.05	0.04	0.07	0.02	0.02	0.01	0.05
TiO₂	0.06	0.05	0.06	0.05	0.06	0.08	0.18	0.12	0.20	0.66	0.33	0.53	0.23	0.51
Al₂O₃	55.23	59.74	25.79	49.98	60.57	57.98	40.40	52.99	54.93	55.82	31.87	16.25	36.41	19.45
Cr₂O₃	10.89	7.80	39.63	16.29	6.17	9.06	27.00	15.44	9.27	9.91	34.58	49.46	30.31	46.18
Fe₂O₃	3.45	1.53	5.78	3.75	3.03	2.34	3.14	1.31	4.59	1.57	4.48	6.03	4.15	5.77
FeO	7.43	9.38	12.17	9.94	7.59	8.55	10.87	9.19	13.61	10.20	13.53	14.14	12.91	14.03
MnO	0.12	0.13	0.23	0.18	0.10	0.12	0.21	0.11	0.24	0.08	0.16	0.19	0.13	0.17
NiO	0.39	0.40	0.00	0.36	0.31	0.42	0.26	0.35	0.29	0.36	0.17	0.16	0.22	0.14
MgO	21.12	20.52	15.21	19.15	21.86	20.80	17.55	20.02	17.51	19.90	15.21	13.08	16.11	13.49
CaO	0.03	0.01	n.d	0.01	0.00	0.01	0.01	0.00	0.00	0.02	<0.02	<0.02	<0.02	<0.02
Na₂O	0.01	0.01	n.d	0.01	0.01	0.03	0.02	n.d	n.d	n.d	n.d	n.d	n.d	n.d
K₂O	0.01	0.01	n.d	0.01	0.00	0.01	0.01	n.d	n.d	n.d	n.d	n.d	n.d	n.d
Total	98.79	99.59	98.93	99.77	99.79	99.45	99.66	99.58	100.68	98.59	100.35	99.86	100.48	99.79
Formulae based on 32 Oxygens & 16 Tetrahedral Cations														
Si⁴⁺	0.010	0.005	0.015	0.012	0.016	0.010	0.007	0.010	0.009	0.015	0.005	0.006	0.001	0.011
Ti⁴⁺	0.009	0.008	0.011	0.008	0.010	0.012	0.029	0.019	0.032	0.105	0.058	0.101	0.04	0.095
Al³⁺	13.633	14.482	7.349	12.616	14.519	14.132	10.657	13.197	13.680	13.902	8.768	4.846	9.784	5.713
Cr³⁺	1.804	1.268	7.575	2.759	0.992	1.481	4.777	2.578	1.549	1.654	6.381	9.897	5.464	9.099
Fe³⁺	0.544	0.237	1.052	0.604	0.464	0.364	0.529	0.208	0.730	0.250	0.787	1.149	0.712	1.081
Fe²⁺	1.301	1.612	2.460	1.780	1.292	1.479	2.035	1.624	2.405	1.802	2.641	2.992	2.461	2.925
Mn²⁺	0.022	0.022	0.047	0.032	0.018	0.021	0.040	0.020	0.044	0.014	0.032	0.04	0.026	0.035
Ni²⁺	0.066	0.066	0.000	0.062	0.051	0.070	0.046	0.057	0.048	0.060	0.033	0.032	0.04	0.028
Mg²⁺	6.592	6.289	5.480	6.112	6.624	6.411	5.855	6.304	5.512	6.264	5.295	4.935	5.474	5.011
Ca²⁺	0.007	0.002	0.000	0.002	0.001	0.002	0.001	0.000	0.000	0.005	0.000	0.000	0.000	0.000
Na⁺	0.002	0.004	0.000	0.003	0.004	0.012	0.007	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K⁺	0.002	0.001	0.000	0.002	0.001	0.002	0.002	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣCations	24	24	24	24	24	24	24	24	24	24	24	24	24	24
Calculated Site Occupancy														
Total_{iv}	16	16	16	16	16	16	16	16	16	16	16	16	16	16
Total_{vi}	8	8	8	8	8	8	8	8	8	8	8	8	8	8
Σ Cations	24	24	24	24	24	24	24	24	24	24	24	24	24	24
Cr#	11.68	8.05	50.76	17.94	6.39	9.49	30.95	16.34	10.17	10.63	42.12	67.13	35.83	61.43

Selected Analyses of Minerals in Scottish Peridotites

Table 2. Selected orthopyroxene analyses.														
	C Fell C	CnB 51	RNB 5	Hex 1	HN 31	RP 77	Rsk 1	STP5	STP1	FD124	FD127	LR80	LR81	
SiO ₂	52.91	55.31	56.20	54.16	54.59	54.64	55.90	55.23	55.96	54.76	54.30	57.06	56.66	
TiO ₂	0.18	0.10	0.02	0.09	0.06	0.08	0.07	0.14	0.11	0.16	0.00	0.09	0.09	
Al ₂ O ₃	5.38	3.74	2.01	4.08	5.11	4.30	2.27	4.41	3.85	4.30	4.37	2.16	2.65	
Cr ₂ O ₃	0.63	0.25	0.29	0.41	0.21	0.31	0.37	0.35	0.40	0.41	0.29	0.42	0.38	
FeO	5.14	6.66	6.18	7.65	6.06	6.45	5.55	5.35	6.05	6.90	8.39	6.34	6.56	
MnO	0.13	0.17	0.20	0.19	0.14	0.16	0.14	0.15	0.14	0.14	0.23	0.18	0.18	
NiO	0.10	0.09	n.d	0.11	0.07	0.10	0.09	0.00	0.13	n.d	n.d	0.06	0.08	
MgO	27.22	33.34	33.93	32.54	32.86	32.99	34.69	33.09	32.61	32.11	31.24	33.77	33.62	
CaO	7.49	0.34	0.81	0.70	0.95	0.68	0.37	0.79	0.58	0.75	0.82	0.45	0.48	
Na ₂ O	0.55	0.08	0.12	0.11	0.08	0.14	0.04	0.13	0.11	0.27	0.39	0.10	0.09	
K ₂ O	0.01	0.00	0.02	0.01	0.00	0.01	0.01	0.01	n.d.	n.d.	n.d.	n.d.	n.d.	
Total	99.73	100.07	99.78	100.06	100.13	99.86	99.50	99.65	99.93	99.81	100.03	100.60	100.76	
Formula based on 6 Oxygens														
Si ⁴⁺	1.868	1.912	1.947	1.887	1.884	1.895	1.935	1.907	1.930	1.903	1.898	1.959	1.943	
Ti ⁴⁺	0.005	0.003	0.001	0.002	0.002	0.002	0.002	0.004	0.003	0.004	0.000	0.002	0.002	
Al ³⁺	0.224	0.152	0.082	0.168	0.208	0.176	0.093	0.179	0.156	0.176	0.180	0.088	0.107	
Cr ³⁺	0.018	0.007	0.008	0.011	0.006	0.009	0.010	0.010	0.011	0.011	0.008	0.012	0.010	
Fe ²⁺	0.152	0.193	0.179	0.223	0.175	0.187	0.161	0.154	0.174	0.201	0.245	0.182	0.188	
Mn ²⁺	0.004	0.005	0.006	0.006	0.004	0.005	0.004	0.004	0.004	0.004	0.007	0.006	0.005	
Ni ²⁺	0.003	0.002	0.000	0.003	0.002	0.003	0.003	0.000	0.004	0.000	0.000	0.002	0.002	
Mg ²⁺	1.432	1.718	1.751	1.690	1.690	1.705	1.789	1.703	1.677	1.663	1.627	1.729	1.719	
Ca ²⁺	0.283	0.013	0.030	0.026	0.035	0.025	0.014	0.029	0.021	0.028	0.031	0.017	0.018	
Na ⁺	0.037	0.005	0.008	0.008	0.005	0.010	0.003	0.009	0.007	0.018	0.026	0.006	0.006	
K ⁺	0.000	0.000	0.001	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	
ΣCations	4	4	4	4	4	4	4	4	4	4	4	4	4	
End-Members														
En	76.5	89.1	89.1	86.9	88.8	88.7	90.9	90.1	89.6	87.9	85.5	89.4	89.1	
Fs	8.3	10.2	9.4	11.8	9.4	10.0	8.4	8.4	9.3	10.6	12.9	9.7	10.0	
Wo	15.1	0.7	1.5	1.3	1.8	1.3	0.7	1.5	1.1	1.5	1.6	0.9	0.9	

Selected Analyses of Minerals in Scottish Peridotites

Table 2 cont. Selected clinopyroxene analyses.													
	C Fell C	CnB 51	RNB 5	Hex 1	HN 31	RP 77	RP289	Rsk 1	STP5	FD124	FD127	LR80	LR81
SiO ₂	51.35	52.08	52.44	51.58	51.36	51.84	51.36	52.02	52.36	52.14	52.01	54.55	53.78
TiO ₂	0.38	0.70	0.20	0.28	0.30	0.50	0.76	0.31	0.58	0.67	0.72	0.34	0.42
Al ₂ O ₃	6.03	6.22	3.92	5.79	6.47	6.87	5.35	3.41	6.71	6.92	7.02	4.22	4.71
Cr ₂ O ₃	0.86	0.56	1.31	0.87	0.34	0.68	1.30	1.07	0.71	0.74	0.72	1.09	0.86
FeO	2.62	2.37	2.96	3.76	3.25	2.83	4.65	1.94	2.61	3.08	3.46	3.10	3.08
MnO	0.09	0.08	0.11	0.12	0.09	0.10	0.19	0.08	0.09	0.00	0.08	0.08	0.08
NiO	0.05	0.05	n.d	0.06	0.05	0.06	0.04	0.05	0.00	0.00	0.18	0.02	0.04
MgO	15.49	14.57	16.58	15.46	16.47	15.02	15.80	16.40	15.44	15.21	15.32	14.69	14.63
CaO	21.36	20.98	21.00	20.09	20.69	19.98	19.75	23.11	19.65	20.38	19.75	19.80	20.45
Na ₂ O	1.43	1.98	1.06	1.59	0.96	1.93	0.90	0.77	1.84	1.54	1.33	2.56	2.32
K ₂ O	0.00	0.01	0.01	0.00	0.01	0.01	n.d.	0.01	0.01	n.d.	n.d.	n.d.	n.d.
Total	99.65	99.59	99.59	99.60	99.99	99.82	100.09	99.16	100.00	100.67	100.43	100.45	100.37
Formula based on 6 Oxygens													
Si ⁴⁺	1.872	1.893	1.913	1.884	1.862	1.879	1.873	1.909	1.888	1.874	1.871	1.959	1.934
Ti ⁴⁺	0.010	0.019	0.005	0.008	0.008	0.014	0.021	0.009	0.016	0.018	0.019	0.009	0.011
Al ³⁺	0.259	0.266	0.169	0.249	0.276	0.294	0.230	0.148	0.285	0.293	0.298	0.178	0.200
Cr ³⁺	0.025	0.016	0.038	0.025	0.010	0.020	0.037	0.031	0.020	0.021	0.020	0.031	0.024
Fe ²⁺	0.080	0.072	0.090	0.115	0.099	0.086	0.142	0.059	0.079	0.093	0.104	0.093	0.093
Mn ²⁺	0.003	0.002	0.003	0.004	0.003	0.003	0.006	0.002	0.003	0.000	0.002	0.002	0.002
Ni ²⁺	0.001	0.002	0.000	0.002	0.002	0.002	0.001	0.001	0.000	0.000	0.005	0.001	0.001
Mg ²⁺	0.841	0.789	0.901	0.842	0.890	0.811	0.859	0.897	0.830	0.815	0.821	0.786	0.784
Ca ²⁺	0.834	0.817	0.821	0.786	0.804	0.776	0.772	0.909	0.759	0.785	0.761	0.762	0.788
Na ⁺	0.101	0.140	0.075	0.113	0.067	0.136	0.064	0.055	0.129	0.107	0.093	0.178	0.162
K ⁺	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣCations	4	4	4	4	4	4	4	4	4	4	4	4	4
End-Members													
En	47.8	47.0	49.6	48.2	49.6	48.4	48.5	48.0	49.7	48.1	48.7	47.9	47.1
Fs	4.7	4.4	5.2	6.8	5.6	5.3	8.0	3.3	4.9	5.5	6.2	5.7	5.6
Wo	47.4	48.6	45.2	45.0	44.8	46.3	43.5	48.7	45.5	46.4	45.1	46.6	47.3

Table 3. Feldspar analyses.

	LR80	LR81
SiO₂	65.87	65.86
TiO₂	0.03	0.04
Al₂O₃	21.20	21.46
Cr₂O₃	0.00	0.00
FeO	0.16	0.23
MnO	0.00	0.00
NiO	0.00	0.00
MgO	0.09	0.04
CaO	2.25	2.39
Na₂O	9.38	9.64
K₂O	1.33	0.94
Total	100.31	100.60
Formula based on 32 Oxygens		
Si⁴⁺	11.613	11.566
Ti⁴⁺	0.004	0.005
Al³⁺	4.405	4.442
Cr³⁺	0.000	0.000
Fe²⁺	0.024	0.034
Mn²⁺	0.000	0.000
Ni²⁺	0.000	0.000
Mg²⁺	0.024	0.010
Ca²⁺	0.425	0.450
Na⁺	3.206	3.282
K⁺	0.299	0.211
ΣCations	20	20
End-Members		
Ab	81.6	83.3
An	10.8	11.4
Or	7.6	5.3

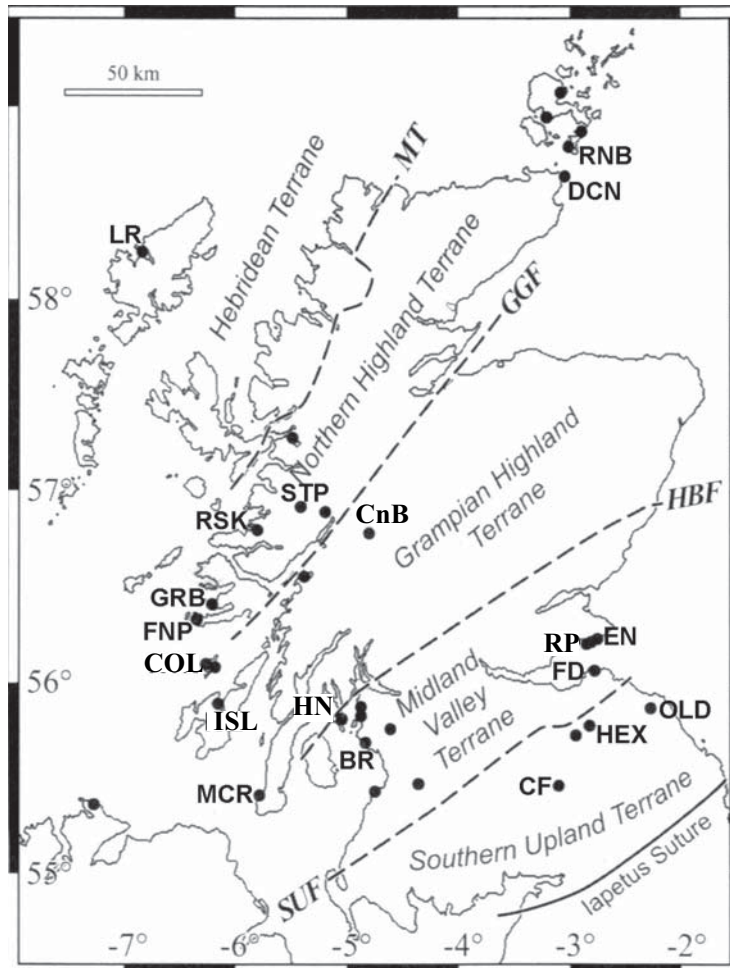


Figure 1.

Figure 2.

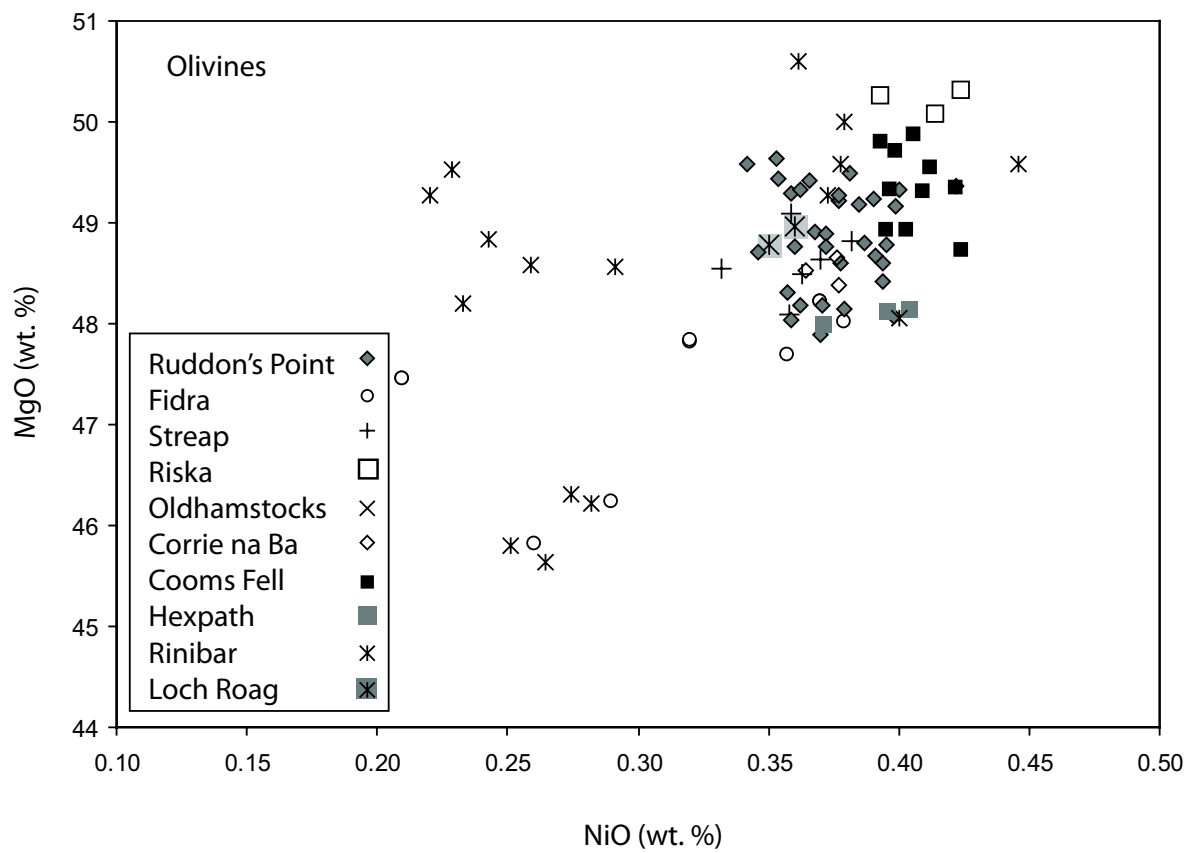
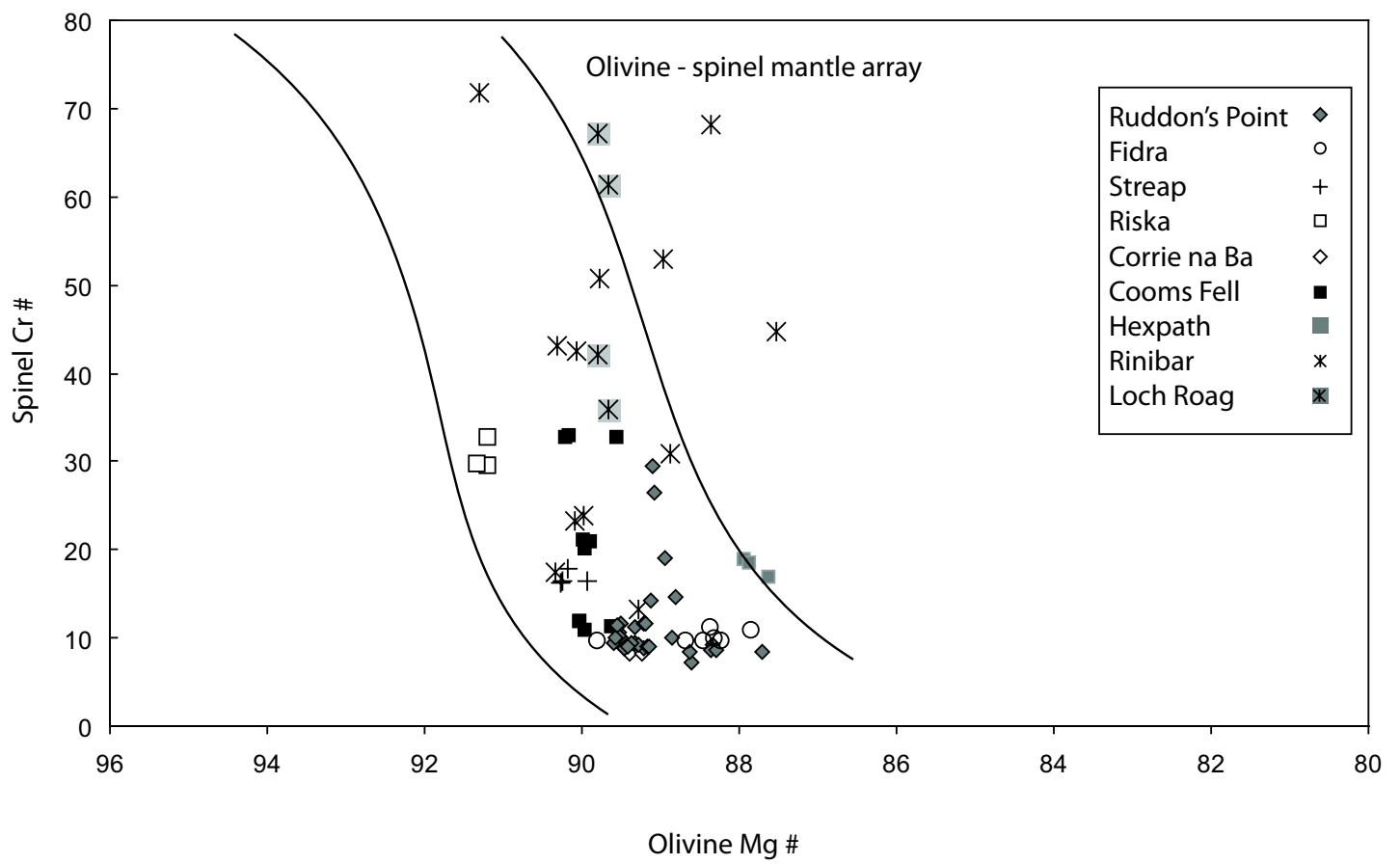


Figure 3.



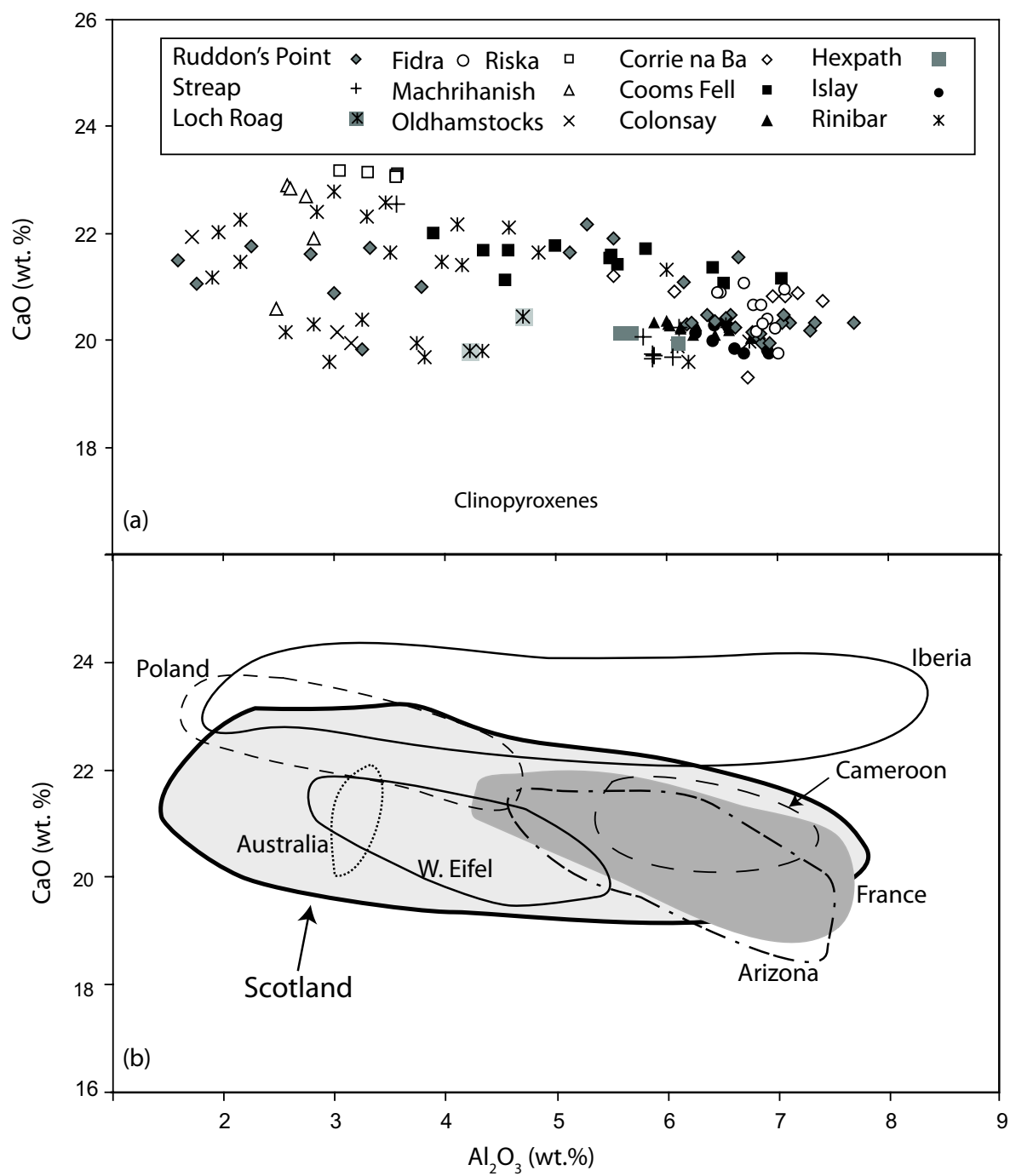
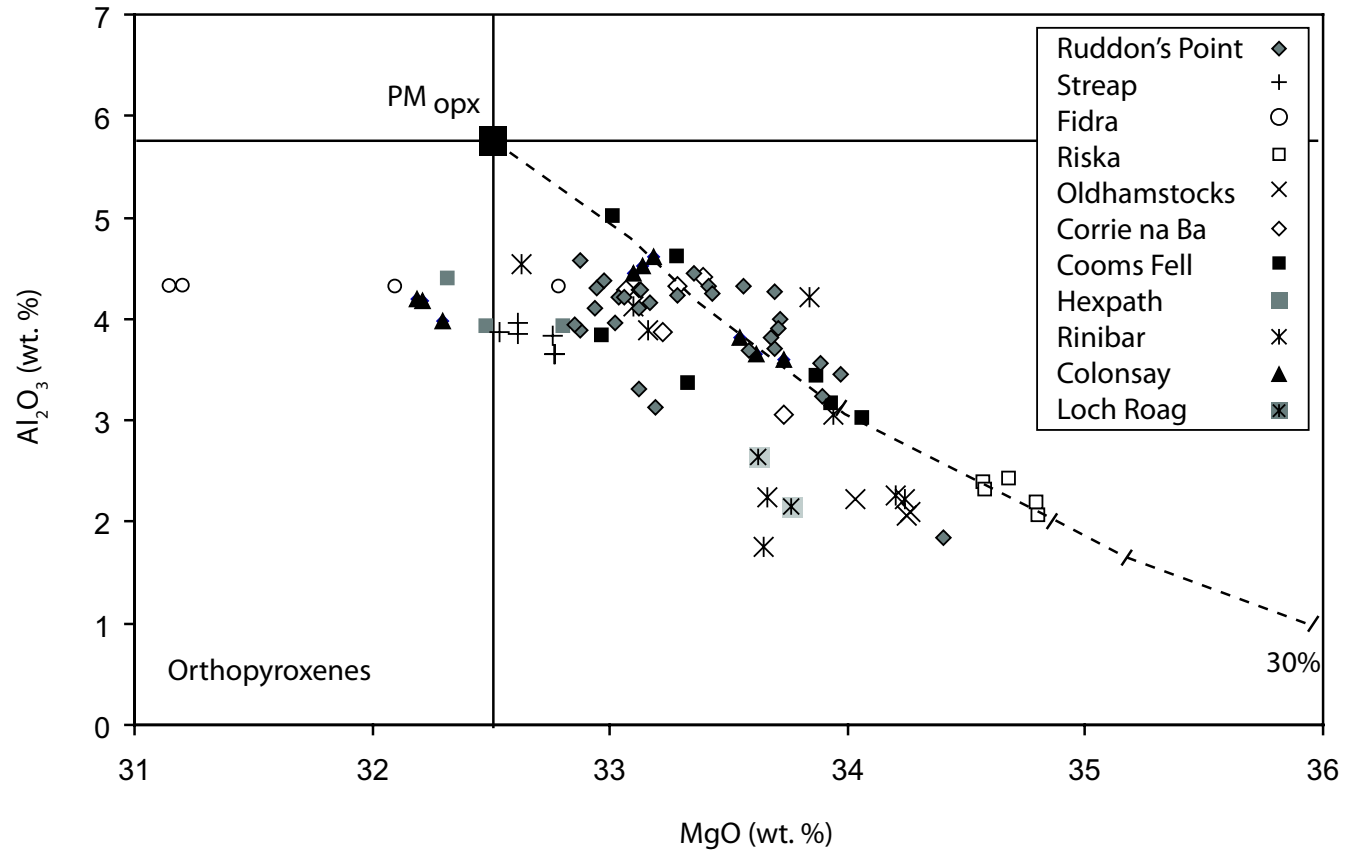


Figure 5.

Figure 6.



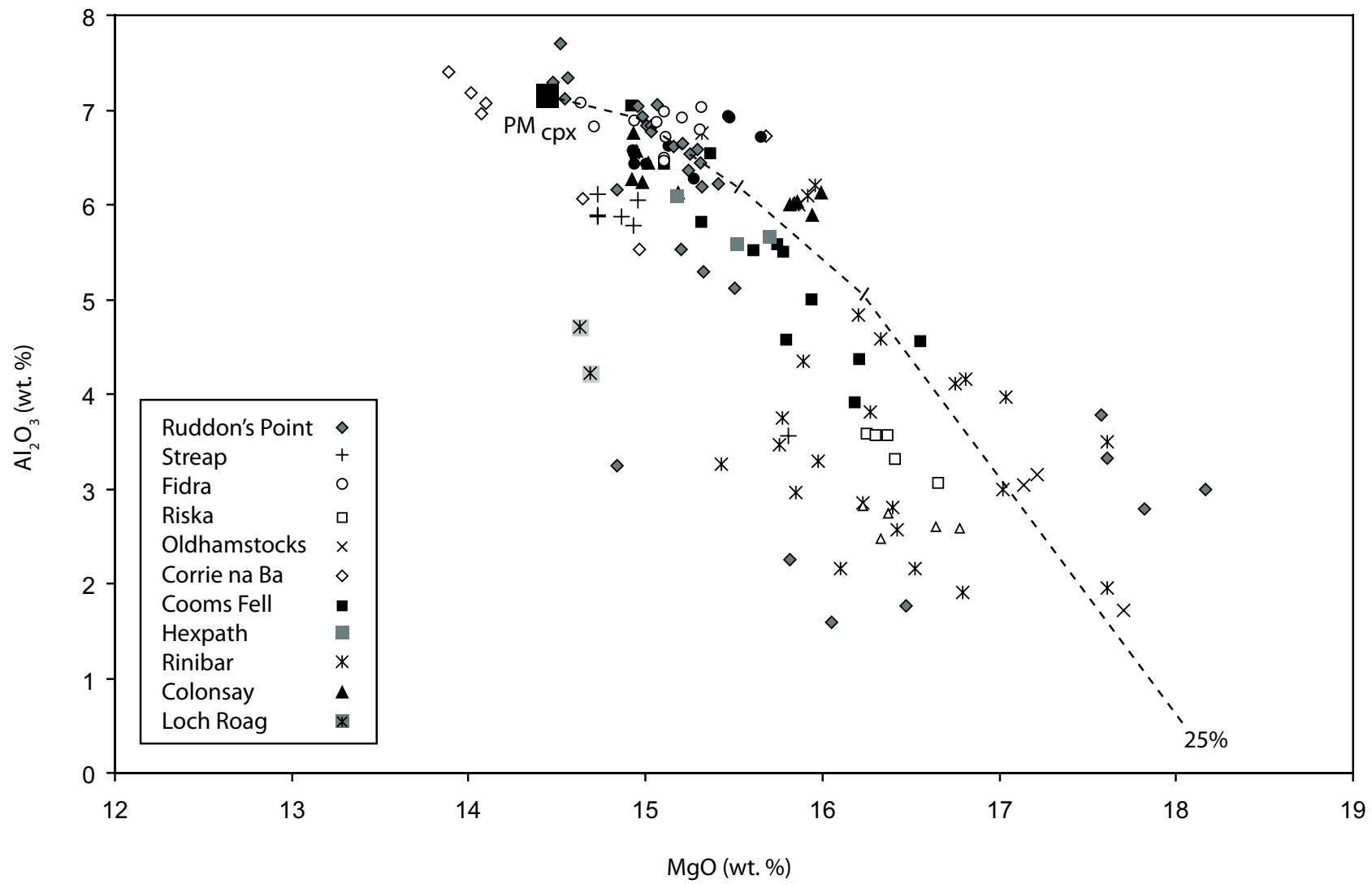
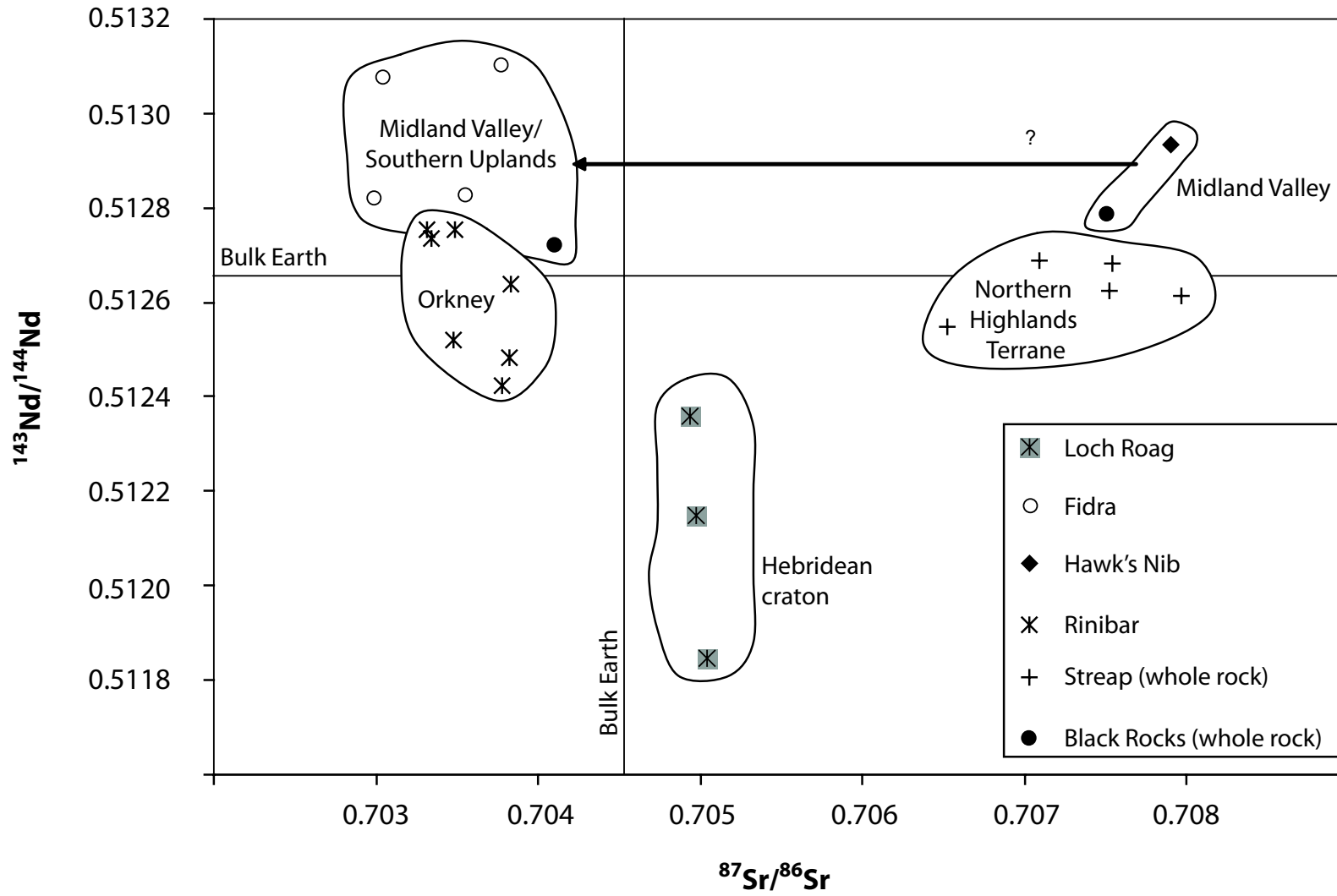


Figure 7.

Figure 8.



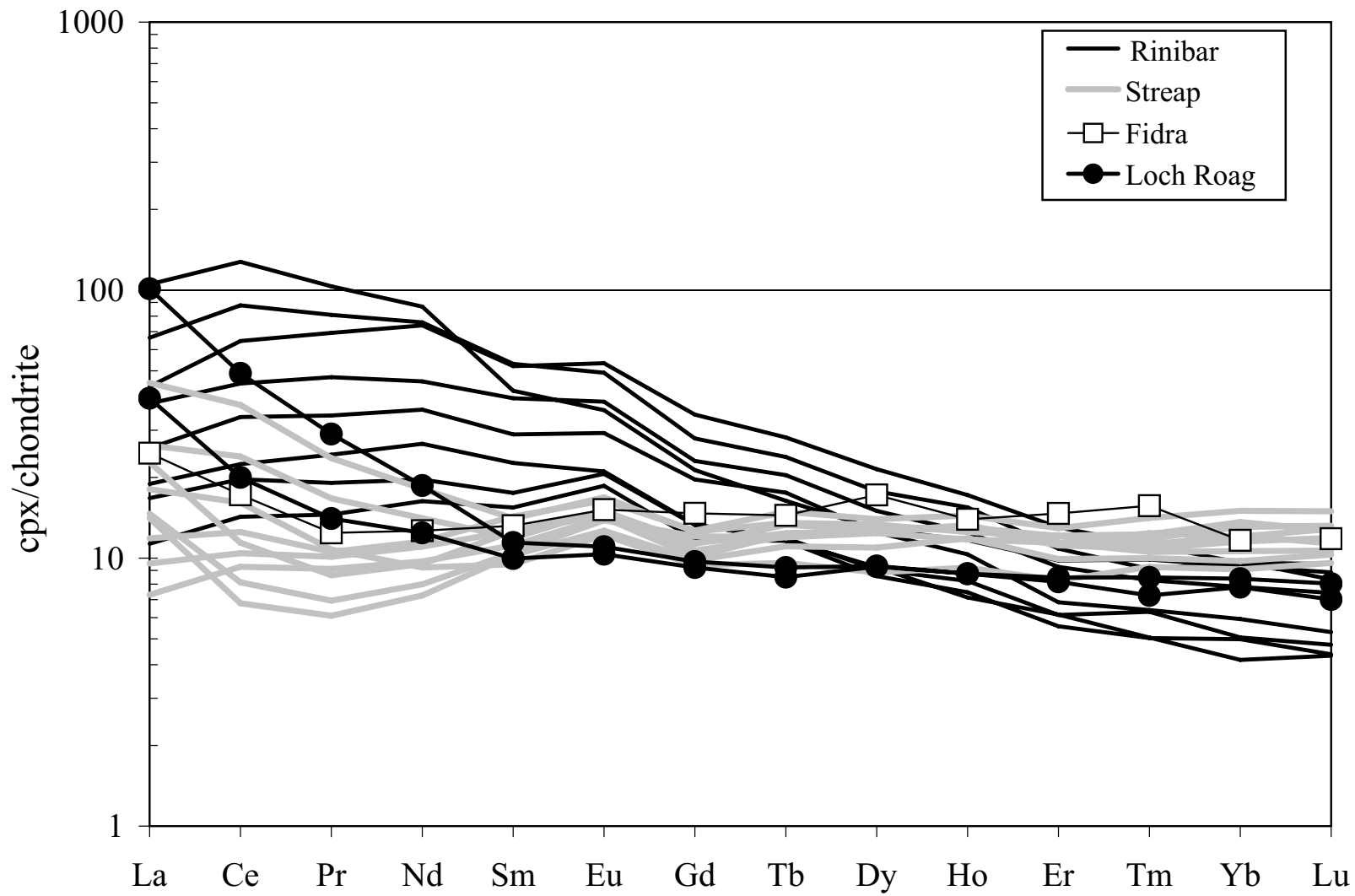


Figure 9.