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1 Mantle sources and magma evolution beneath the Cameroon Volcanic Line: Geochemistry of mafic
2 rocks from the Bamenda Mountains (NW Cameroon).

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Abstract

We report the mineralogy, geochemistry and geochronology of the mafic rocks from the Bamenda Mountains, part of the the Cameroon Volcanic Line (CVL), in order to discuss the origin and evolution of the magmas in this part of the CVL.

Mafic rocks in the Bamenda Mountains are basanites, basalts, hawaiiites and mugearites with an alkaline affinity. K-Ar ages have been obtained on 10 samples and range from 17.6 Myr to present.

Trace element and isotopic compositions (Sr-Nd-Pb) show that some samples among the oldest are slightly contaminated by a crustal component with high La/Nb and ⁸⁷Sr/⁸⁶Sr ratios and low Pb isotopic ratios.

The mafic rocks strongly resemble OIB in their trace element compositions. Some samples possess a positive Sr and Eu anomaly which cannot be explained by a process of plagioclase accumulation.

These anomalies are also observed in some pyroxenites found as xenoliths in the Adamawa volcanic province further north.

Furthermore, non-contaminated samples have high Pb isotopic ratios and point towards an HIMU component similar to the St Helena mantle plume. We propose that the Bamenda mafic magmas with positive Sr and Eu anomalies were formed by hybridation of asthenospheric melts with melts formed by the partial melting of pyroxenites. Samples without these anomalies result from the hybridation of

35 the same asthenospheric melts with melts coming from the metasomatized, amphibole-bearing,
36 lithospheric mantle.

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38 **Keywords:** Cameroon Volcanic Line; Bamenda Mountains; mafic lavas; crustal contamination;
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39 mantle sources; pyroxenites.

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1. Introduction

Intraplate mafic magmatism generally has a wide chemical diversity especially in its trace element and radiogenic isotope compositions. The variations are considered to arise from chemically heterogeneous sources deep in the mantle (e.g., Stracke et al., 2005 and references therein), but the interaction between upwelling plume and the lithospheric mantle or the crust is another possibility to produce chemical diversities (e.g., Class and Goldstein, 1997; Bourdon et al., 1998; Class et al., 1998; Claude-Ivanaj et al., 1998; MacDonald et al., 2001; Lundstrom et al., 2003; Rankenburg et al., 2005; Yokoyama et al., 2007), making much more complicated to fully understand magma generation processes in continental intraplate tectonic settings.

The Cameroon Volcanic Line (CVL) is a chain of 12 Cenozoic volcanic massifs running for approximately 1600 km from the island of Annobon in the Gulf of Guinea to Lake Chad (Fig. 1a). It can be divided into three zones: the oceanic sector (Annobon, Sao Tome and Principe), the continent/ocean boundary (c.o.b.: Bioko, Etna and Mt. Cameroon) and the continental sector. The continental sector is marked by a trend of large massifs including the Manengouba, Bambouto, Bamenda and Oku mountains (Fig. 1b). Fitton and Dunlop (1985) found geochemical similarity in trace elements and Sr isotopes for basalts from both the oceanic and continental sectors, and suggested that these magmas are derived from sub-lithospheric depths without further interaction with the overlying lithosphere. Halliday et al. (1988, 1990), however, reported anomalously high $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (up to 20.5) for the c.o.b. volcanoes, compared with relatively lower (19-20) $^{206}\text{Pb}/^{204}\text{Pb}$ ratios for the oceanic and continental sector volcanoes. With the combination of other tracers such as Sr, Nd and O isotopes, they demonstrated that such a high $^{206}\text{Pb}/^{204}\text{Pb}$ signature was created by the remelting

62 and U/Pb fractionation during melt migration in the St. Helena fossil plume head as it cooled after
63 emplacement 125 Myr ago. The study concluded that Cameroon line magmas are currently derived
64 from a zone in the upper portions of the fossil plume in the lithospheric mantle. This model implies
65 that the observed Pb isotope anomaly did not derive from metasomatized lithosphere. Recent research
66 of helium isotopes for the CVL showed MORB-like $^3\text{He}/^4\text{He}$ ratios for the oceanic and continental
67 sectors and HIMU-OIB-like ratios for the c.o.b. (Aka et al., 2004), which can consistently be
68 explained by the fossil plume remelting model.

69 The Bamenda mountains are characterized by the predominance of felsic lavas over mafic ones
70 (Kamgang et al., 2007; 2008; 2010). We report major elements, trace elements, ages and Sr-Nd-Pb
71 isotopes of the Bamenda lavas in order to give further constraints on the magma generation.

73 2. Geological setting

74 The Bamenda Mountains are located between Lat. $5^{\circ}40'$ - $6^{\circ}10'N$ and Long. $10^{\circ}00'$ - $10^{\circ}30'E$ and
75 culminate at 2,621m. They constitute with the Bambouto Mountains to the SW and the Oku massif to
76 the NE, the three central volcanic massifs belonging to the volcanic centres of the Western Cameroon
77 Highlands (WCH, Fig. 1c). Their basement consists of Pan-African granitoids (Toteu *et al.*, 2001;
78 Nzolang *et al.*, 2003; Nzolang, 2005). The petrological and geochemical features of Bambouto
79 Mountains and Oku massif are more or less established (Marzoli *et al.*, 1999, 2000). The common
80 features of these volcanic massifs are the predominance of felsic lavas over the mafic ones.

81 The Bamenda Mountains (Kamgang et al., 2007, 2008, 2010) are one of the most important volcanoes
82 of the Cameroon Line in North-West Cameroon. They are mainly made up of mafic and felsic lavas
83 with very small amounts of intermediate terms (benmoreite). Mafic lavas mostly consist of basanites,
84 alkaline basalts, hawaiites and mugearites.

86 3. Analytical procedures

87 Mineral analysis of Bamenda mafic lavas was performed using a Cameca Camebax electron
88 microprobe at Nancy I, France (15 kV accelerating potential, 10-12 nA current and 10 s counting time
89 per element). The main mineralogical data are given in Table 1.

90 38 samples have been analysed for major and trace elements. Samples were crushed in a steel
91 jaw crusher and reduced to fine powder in agate mortars. Powders were analysed for major oxides and
92 Sc by inductively coupled plasma-atomic emission spectrometry (ICP-AES), and for other trace
93 elements by inductively coupled plasma-mass spectrometry (ICP-MS) at the Centre de Recherches
94 Pétrographiques et Géochimiques (CRPG, Nancy, France). Precision is 0.5 % for major elements, and
95 variable for trace element contents: 2-5 % in the range 50-150 ppm, 2-10 % ppm in the range 10-50
96 ppm, 5-25 % in the range 0-10 ppm. REEs are measured with a general precision of 5 %, and a
97 precision always better than 10 % (Morel, CRPG, personal communication 1997). More details about
98 the analytical procedures can be found in Carignan *et al.* (2001).

99 Sr and Nd isotopic compositions were measured at Clermont-Ferrand, France. Sr was isolated by
100 standard cation exchange and loaded onto a single Ta filament with phosphoric acid. Nd was isolated
101 by cation exchange, then purified by Di-(2-ethylhexyl) phosphoric acid (HDEHP) Teflon substrate
102 column and loaded onto Ta filaments of a triple Ta-Re-Ta filament assembly. Sr and Nd isotopic ratios
103 were measured on a VG 54E mass spectrometer, in double collection mode. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are
104 normalized to $^{86}\text{Sr}/^{87}\text{Sr} = 0.1194$. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are normalized to $^{143}\text{Nd}/^{144}\text{Nd} = 0.7219$. Over
105 the period of measurements, replicate analyses of standards gave, for National Bureau of Standards
106 (NBS) 987, $^{87}\text{Sr}/^{86}\text{Sr} = 0.71024 \pm 4 (2\sigma)$ and, for La Jolla, $^{143}\text{Nd}/^{144}\text{Nd} = 0.51184 \pm 3 (2\sigma)$. Routine
107 analytical blanks at Clermont-Ferrand were $<0.4\%$ for Nd and $<0.03\%$ for Sr. Pb isotope ratios were
108 measured either on a Cameca TSN 206 SA mass spectrometer or a VG 54E (Clermont-Ferrand) in
109 single collection mode.

110 Ten K–Ar radiometric ages for mafic lavas were determined from whole-rocks at the CRPG in
111 Nancy following the procedure described in Zimmermann *et al.* (1985) which is summarized below or
112 at the University of Queensland in Australia. Potassium content is determined by flame photometry

113 with a lithium internal standard. Argon is extracted in a heat-resistant glass vacuum apparatus and
114 determined by isotope dilution (^{38}Ar as tracer) using a MS20 mass-spectrometer. All samples are
115 measured in the static mode. Age calculations were carried out using the constants recommended by
116 Steiger and Jäger (1977). Errors were calculated with the error equation described in Defant et al.
117 (1992), which is derived from Mahood and Drake (1982). The dated samples fulfil the following
118 criteria: (a) absence of secondary K-bearing minerals (phyllosilicates, zeolites) in thin section, (b) loss
119 on ignition values lower than 3%.

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4. Nomenclature and petrography

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All major-oxides have been normalized to 100% on an anhydrous basis. Data are plotted on the total alkali vs. silica diagram (Le Bas et al., 1986) with subalkaline-alkaline dividing line of Miyashiro (1978) for the purpose of classification (Fig. 2). According to this diagram, all the samples exhibit an alkaline trend. Based on whole rock major element chemical compositions only, the Bamenda samples range in composition from basanite, basalt, hawaiite to mugearite. In the Na_2O vs. K_2O diagram shown in Figure 2 the Bamenda mafic lavas plot mainly in the Na-series field ($\text{Na}_2\text{O}/\text{K}_2\text{O} = 1.54\text{-}4.25$) of Middlemost (1975).

Basanite is the dominant lava type. The basanites range from porphyritic to aphyric in texture. The porphyritic varieties have abundant (10-15%) large (up to 2 mm) euhedral to subhedral clinopyroxene and olivine crystals. The phenocrysts of clinopyroxene show very complex zoning with light pink core and a lilac coloured rim. In some samples, clinopyroxenes with an olive-green to light green core, and a dark brown rim occur in addition to the clinopyroxene described above. Some large olivines show resorption and strain lamellae, and are thus interpreted as xenocrysts. The olivine xenocrysts are strongly altered to red-brown iddingsite or serpentine. They occur in basanite BA10. Subhedral to euhedral titanomagnetite occurs as inclusions in both clinopyroxene and olivine and as euhedral microcrysts in the groundmass. The aphyric basanites display mineralogy similar to porphyritic varieties with rare phenocrysts of clinopyroxenes and ubiquitous microphenocrysts (< 1 mm) of olivine, clinopyroxene, and titanomagnetite.

140 The groundmass is composed of plagioclase, clinopyroxene, olivine, titanomagnetite smaller than 0.05
141 mm occasionally coexisting with interstitial alkali feldspar and nepheline in some samples.

142 Basalts are porphyritic with olivine and clinopyroxene (15-20%) as main phenocrysts (2-3mm);
143 in some samples they coexist with plagioclase phenocrysts. Two types of clinopyroxene phenocrysts
144 have been distinguished: mainly light pink core and lilac coloured rim and few with green core and
145 lilac coloured rim. Microphenocrysts of olivine, pink clinopyroxene and titanomagnetite are embayed
146 in a intergranular matrix made of clinopyroxene, plagioclase, olivine, and titanomagnetite and
147 secondary calcite in some samples.

148 Hawaiites contain abundant (10-15%) plagioclase phenocrysts (laths of 2-3 mm in size), less
149 abundant olivine phenocrysts (5 mm in size) and smaller clinopyroxene phenocrysts sometimes with a
150 light green core. Microphenocrysts include the above phases and titanomagnetite. Some of these lavas
151 contain rounded quartz xenocrysts. The groundmass comprises plagioclase, olivine, clinopyroxene,
152 Fe-Ti oxides and biotite and acicular needles of apatite in some samples.

153 Mugearites range from porphyritic to aphyric in texture. The porphyritic varieties (e.g., BA39)
154 contain phenocrysts of plagioclase accompanied by microphenocrysts of zoned clinopyroxene (light
155 pink core) and more rarely olivine and Fe-Ti oxides. Olivines are mostly altered. The aphyric varieties
156 comprise microphenocrysts of olivine and Fe-Ti oxides. The groundmass of mugearites contains
157 plagioclase, olivine, clinopyroxene, calcite and euhedral microcrystal of apatite. Like others Bamenda
158 mafic lavas, scarce alkali feldspars are also present in the groundmass of the mugearites.

160 5. Mineral chemistry

161 Olivine

162 The range of olivine composition is depicted in Figure 3 (Suppl. data). Olivine composition in
163 basanites range from Fo₆₁ to Fo₈₇ with 0.10-0.59 wt.% CaO and 0.08-0.72 wt.% MnO. In these rocks,
164 olivine phenocrysts are generally more magnesian and exhibit zoning, with Fe enrichment from core
165 (Fo₇₀₋₈₅) to rim (Fo₆₇₋₆₉). Microphenocrysts are more Fe-rich (Fo₆₁₋₆₉). The higher Fo contents of the
166 observed olivine xenocrysts (Fo₈₉₋₉₁) exceed the equilibrium value indicating their incomplete re-

167 equilibration with the host magma. The more primitive character of these xenocrysts, also attested by
168 their higher Ni and lower Mn and Ca contents (Suppl. data) is rather characteristic of mantle-derived
169 olivine peridotite.

170 In the basalts, olivine composition ranges from Fo₅₂ to Fo₈₄ with 0.16-1.16 wt.% CaO and 0.15
171 -0.86 wt.% MnO. In the hawaiites, olivine composition ranges from Fo₅₆ to Fo₆₇ with 0.26-0.47 wt.%
172 CaO and 0.63-0.83 wt.% MnO. In the mugearites, olivine composition ranges from Fo₄₂ to Fo₅₀ with
173 0.34-1.06 wt.% CaO and 0.96-1.38 wt.% MnO.

174 Although there is significant overlap in composition amongst the different groups of lavas,
175 there is a general increase in fayalite component and MnO content with degree of differentiation.

176 Clinopyroxene

177 Chemical compositions of the representative Cpx analyses are shown in a ternary plot of the
178 Wo-En-Fs system (Fig. 3; Suppl. data). All Cpx from different rock-types are Ca-rich and range from
179 diopside to augite in composition.

180 Cpx in basanites ranges from Wo₅₂En₃₃Fs₁₅ to Wo₄₆En₄₁Fs₁₃. TiO₂ (2.1-6.3 wt.%), Al₂O₃ (3.9-
181 10.2 wt.%) and Na₂O (0.4-0.8 wt%) abundances are fairly high and varied and indicate the alkali
182 nature of these pyroxenes (Schweitzer et al., 1979). Cpx in some basanite shows both concentric and
183 sector zoning. Sector-zoned Cpx phenocrysts show differences in TiO₂, Al₂O₃, FeO and MgO contents
184 in different sectors (Fig.4). Green cores noted in some of the larger phenocrysts are Na₂O and FeO rich
185 but low in Al₂O₃, MgO and TiO₂ (Fig. 4). Al/Ti ratios are low in the core (0.04-0.19) and high in the
186 rim (0.27-0.39). The green-core Cpx belong to aluminian diopside according to the IMA classification
187 scheme (Morimoto et al., 1988) and are fassaitic or ferrosalitic. Both types are comparatively Fe rich.

188 In the basalts the green Cpx were not analysed. Cpx range from Wo₄₇En₃₈Fs₁₅ to Wo₄₄En₄₀Fs₁₆.
189 TiO₂ (1.1-3.7 wt.%), Al₂O₃ (1.3-6.5 wt.%) and Na₂O (0.00-0.05 wt.%) abundances are low relative to
190 Cpx from basanites. The phenocrysts are normally zoned with decreasing MgO contents from the core
191 towards the rim (Fig.4). FeO and TiO₂ increase from core to rim.

192 Cpx of hawaiites is mainly diopside and ranges in composition from $Wo_{45}En_{37}Fs_{18}$ to
193 $Wo_{50}En_{36}Fs_{14}$. TiO_2 (1.3-3.6 wt%), Al_2O_3 (2.1-4.5 wt.%) and Na_2O (0.5-1.3 wt.%) abundances are low
194 relative to Cpx from basanites. Their Na_2O content is higher than that of basalts.

195 In mugearites, Cpx is augite and ranges in composition from $Wo_{42}En_{29}Fs_{29}$ to $Wo_{45}En_{36}Fs_{19}$.
196 TiO_2 , Al_2O_3 and Na_2O abundances are 0.6-1.2, 1.5-2.7 and 0.3-0.5 wt. % respectively.

197 Feldspars

198 Plagioclase compositions are illustrated in a ternary plot of the An-Ab-Or system (Fig. 3;
199 Suppl. data). The CaO content (1.1-13.5, 1.8-16.4, 0.9-12.2, and 2.0-11.5 wt.% for basanite, basalt,
200 hawaiite and mugearite respectively) shows a wide range and the anorthite content ranges from 5.1 to
201 68.7 wt.% in basanites, 8.7 to 81.1 wt.% in basalts, 4.6 to 61.3 wt.% in hawaiites and 8.4 to 55.1 wt.%
202 in mugearites. The Al_2O_3 and FeO contents show wide and narrow compositional ranges with 19.2-
203 32.4 wt. % and 0.1-1.0 wt.% , respectively. The major-element variations (in wt. % oxide) along rim to
204 rim traverses across one plagioclase phenocryst from basalt BA26 are shown in Figure 4. The rim is
205 lower in Al_2O_3 and CaO and higher in Na_2O than the core.

206 The combined trend of the basanite, basalt, hawaiite and mugearite feldspars describes a classic
207 feldspar solidus (Tuttle and Bowen, 1958), extending from calcic bytownite ($An_{81}Ab_{19}Or_0$) to sodic
208 sanidine ($An_5Ab_{45}Or_{50}$) (Fig. 3).

209 Fe-Ti oxides

210 Titanomagnetite is the only Fe-Ti (Fe recalculated after Stormer, 1983) phase occurring in the
211 mafic lavas of Bamenda (Suppl. data). The Titanomagnetite has variable ulvöspinel contents (49-87
212 mol.%; 14.5-27.9 wt.% TiO_2). MnO ranges from 0.4 wt.% to 1.0 wt.% and Al_2O_3 from 1.0 wt.% to 7.7
213 wt.%.

214 Feldspathoids

215 Nepheline has been analysed as microlites in the basanites. Its composition (mol. %) varies
216 between $Ne_{75}Ks_{11}Qz_{14}$ and $Ne_{63}Ks_6Qz_{31}$ (Suppl. data). Na_2O content ranges from 13.4 to 16.8 wt. %.
217 K_2O (1.6-2.8 wt. %) and Fe_2O_3 (0.3-0.9 wt.%) contents are low relative to nepheline of the phonolite
218 from Upper Benue valley (4.9 and 1.8 wt.% respectively; Ngounouno et al., 2003).

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220 Mica

221 Phlogopite occurs only in hawaiites. It is characterized by high MgO contents, ranging between
222 17.6 and 20.1 wt.%. The FeO and Al₂O₃ contents vary between 8.1-13.7 wt.% and 11.1-12.3 wt.%
223 respectively. It has low Na₂O (0.9-1.1 wt.%) and CaO (0.03-0.4 wt.%), with TiO₂ contents between
224 3.6 and 5.2 wt.% (Suppl. data).

226 6. Results

227 6.1. Geochemistry

228 6.1.1. Major element geochemistry

229 In this paper we focus on the most mafic lavas of the Bamenda volcanic province. The major
230 elements (Table 8) data show that these rocks have SiO₂ ranging from 41 to 52 wt.%, MgO from 2.5
231 to 11.0 wt.% and CaO from 5.3 to 12.7 wt.%. TiO₂ is high, ranging from 2 to 4.4 wt.%.

232 The Mg-number values [$Mg\# = 100 * MgO / (MgO + FeO)$, atomic ratio, assuming that
233 $Fe_2O_3/FeO = 0.15$] lie between 25 and 62, and mostly correlate with major element contents showing
234 fairly coherent trends (Fig. 5). These values mean that the magmas cannot be primary. For SiO₂,
235 Al₂O₃, Na₂O and K₂O, the correlations are negative and, for CaO, positive. However, some samples
236 with Mg# between 45 to 62 show narrow range of CaO contents (8 – 10 wt.%). TiO₂ increases when
237 Mg# decreases from 55 to 45 and then decreases when Mg# is lower than 45. These variations may
238 correspond to those expected from fractional crystallization process, even though plots are scattered.
239 This scatter is probably partly due to the large range of age of the studied samples (from 28 Ma to
240 present for the whole Bamenda lavas) implying that we are observing different differentiation trends
241 from different parental magmas.

242 6.1.2. Trace element geochemistry

244 Compatible trace element geochemistry

245 Trace element data are listed in Table 2. Sc, Cr, Co, Ni and V behave compatibly in the
246 Bamenda basaltic lavas (decrease with falling Mg#). The less evolved rocks (Mg# > 55) have high Cr
247 and Ni contents ranging up to 556 and 283 ppm respectively, but lower than the values assumed for
248 primary magmas (Ni: 300-400 ppm; Cr: 300-500 ppm; Co: 50-70 ppm; e.g., Frey et al., 1978; Jung
249 and Masberg, 1998). The more evolved samples (Mg# < 55) have low Cr and Ni contents ranging
250 down to 0.70 and 1.54 ppm respectively. In Fig. 6a, b, the rocks define a hyperbolic trend presumably
251 reflecting crystal fractionation with the removal of olivine + pyroxene ± spinel assemblage.

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Incompatible trace element geochemistry

254 Harker variation diagrams of trace elements are plotted vs. SiO₂ (Fig. 6). As a whole, plots are
255 scattered. The best correlation is exemplified by Zr which correlates positively with SiO₂. These trends
256 may be related to the fractional crystallization processes (Fig. 6). Again, the scattering in the diagrams
257 may be due to the presence of several differentiation trends through time.

258 Distinct incompatible trace elements characteristics of these rocks include: large variation of
259 total rare earth elements (ΣREE) contents (135 – 401 ppm) and trace element ratios such as Nb/La (0.9
260 – 1.5) and Ba/Nb (7.3 – 21.7). Several other key parameters are: Ba/Nb = 7.3 – 21.7, Ce/Pb = 20.0 –
261 42.0, Nb/U = 33.2 – 81.3, Th/La = 0.06 – 0.13 and Zr/Nb = 2.8 – 6.1.

262 Chondrite-normalized REE patterns of the Bamenda samples (Fig. 7a-d) exhibit light REE
263 (LREE) enrichment relative to heavy REE (HREE) with (La/Yb)_n (where the subscript n means
264 chondrite-normalized) ranging from 10.7 to 22.8. All the analysed samples show positive europium
265 anomalies (Eu/Eu* = 1.1 – 1.4, where Eu is normalized Eu and Eu* is Eu interpolated between
266 normalized Sm and Gd: $Eu^* = (Sm_n \times Nd_n)^{1/2}$). In the absence of any cumulus textures and positive
267 correlation between Al₂O₃ and Eu anomaly, we argue that the positive Eu anomalies in these mafic
268 lavas are an original geochemical feature of the melts rather than a result of plagioclase accumulation.
269 Chondrite-normalized Tb/Yb values (1.8-3.0) are roughly constant among the Bamenda samples, and
270 overlap values for Mount Cameroon (Yokoyama et al., 2007), Ngaoundéré volcanism (Nkouandou et
271 al., 2008) and alkali lavas of Nyos (Aka et al., 2004). This ratio is a good indicator of residual garnet

272 in the mantle source of the basalts, thus suggesting that Bamenda primary magmas have been formed
273 by melting in the presence of garnet. REE patterns of N-MORB, E-MORB and OIB (Sun and
274 McDonough, 1989) are also illustrated in Fig. 7. All Bamenda samples are above those of N-MORB
275 and E-MORB and strongly resemble that of OIB.

276 Incompatible trace elements are very complex as can be seen in primitive mantle normalized
277 incompatible trace elements diagrams (Fig. 8a, b, c, d). All the Bamenda mafic lavas show high
278 concentrations of highly incompatible large ion elements (Ba, Rb, Th, U) relative to moderately
279 incompatible elements (e.g., Ti and HREE). The Bamenda mafic lavas have positive Nb and Ta
280 anomalies and K, P and Ti negative anomalies. Basanites, basalts and hawaiites show positive Sr
281 anomalies whereas mugearite shows negative anomalies. Mantle-normalized Ce/Pb ratios in Bamenda
282 mafic lavas are greater than 1, which is typical of plume-derived magmas (e.g., Tian et al., 2010).

284 6.2. Geochronology

285 The K-Ar data are presented in the Table 3. The ages range from 17.6 Ma to Present. Even if these
286 ages are not representative of the whole volcanic area, they show that mafic volcanism exists over a
287 long period of time and is partly coeval with the felsic volcanism (from 27 to 12 Ma, Kamgang et al.,
288 2007). However, very recent ages obtained on some mafic rocks have not been found among the felsic
289 ones. The age of the mafic volcanism of the Bamenda area is consistent with ages measured for the
290 volcanism of the other volcanic provinces from the Oku massif to the North to the still active Mount
291 Cameroon to the South (Déruelle et al., 2007). Consequently, volcanism in the Bamenda Mountains
292 exists over a more than 25 Myr period, with no displacement of the volcanic centers through time.

294 6.3. Sr, Nd and Pb isotope geochemistry

295 The results of Sr, Nd, Pb isotope analyses are given in Table 4. The effects of age correction for
296 radioactive decay are small for our samples because Rb/Sr and Sm/Nd are low. Almost all of the
297 samples display positive ϵ Nd values ranging from +1.85 to +5.07, typical of mantle-derived magmas
298 (Table 4 and Fig. 9a), the only exception being sample BA60, which shows ϵ Nd = -0.40.

299 As shown in a diagram of ϵ Nd vs. $^{87}\text{Sr}/^{86}\text{Sr}$ (Fig. 9a), the Bamenda mafic lavas plot in the “depleted
300 field” relative to bulk silicate Earth (BSE). The higher $^{143}\text{Nd}/^{144}\text{Nd}$ than Bulk Earth in all Bamenda
301 mafic lavas (except BA60) implies a time-integrated depletion of the source of the magmas. Generally,
302 this trend is similar to trends of other mafic alkaline rocks from the CVL (Halliday et al., 1988, 1990;
303 Lee et al., 1994; Marzoli et al., 2000; Rankenburg et al., 2005; Yokoyama et al., 2007; Nkouandou et
304 al., 2008; Nkouathio et al., 2008). These ϵ Nd values are less variable than those of xenoliths of CVL (-
305 1.4 to +17; Lee et al., 1996) but are in the range of clinopyroxene and plagioclase megacrysts (+3.0 to
306 6.4; Lee et al., 1996; Rankenburg et al., 2005). Some samples of Bamenda mafic lavas with ϵ Nd > 4.0
307 and $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7031-0.7033 fall into the field of FOZO (focal zone) (Hart et al., 1992) recently
308 redefined by Stracke et al. (2005), or “C” (Hanan and Graham, 1996). This field is similar or slightly
309 more radiogenic in its Sr composition than HIMU mantle (HIMU = high μ ; $\mu = ^{238}\text{U}/^{204}\text{Pb}$; e.g.,
310 Zindler and Hart, 1986; Hofmann, 1997). Sample BA60 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7042$ and $^{143}\text{Nd}/^{144}\text{Nd} =$
311 0.51261) plots towards the field of enriched mantle. Its composition is similar to EM1 (EM = enriched
312 mantle; Zindler and Hart, 1986). Some data of Bambouto (Lee et al., 1994; Marzoli et al., 2000) and
313 from Oku (Marzoli et al., 2000) indicate minor contribution of EM1.

314 These isotopic data overlap with the isotopic composition of OIB, and plot around or slightly
315 above the low Nd (LoNd) array (Hart, 1984) in the $^{206}\text{Pb}/^{204}\text{Pb}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ plot (Fig. 9b) except two
316 samples. Samples BA60 and BA41 plot below the LoNd array whereas sample BA60 plots around
317 EM1. $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic ratio is positively correlated with $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 9c). In the $^{206}\text{Pb}/^{204}\text{Pb}$
318 vs $^{207}\text{Pb}/^{204}\text{Pb}$ correlation plots (Fig. 9 d), six samples plot below the LoNd array and around the North
319 Hemisphere Reference Line (NHRL) (Hart, 1984); three samples plot below the NHRL. All the
320 samples show a linear trend between HIMU and EM1 components.

322 7. Discussion

323 7.1. Crustal contamination

324 Continental alkaline basalts are often enriched in trace-elements, with different isotopic ratios relative
325 to their oceanic counterparts and the debate continues as to the importance of processes operating at

326 crustal levels. Given the continental environment in which the alkaline lavas of Bamenda erupted, it is
327 essential to evaluate the possible effects of crustal contamination which may have affected the
328 ascending magmas. Such crustal contamination processes associated with the differentiation of the
329 magmas (AFC process) have already been observed in the other volcanic massifs of the CVL
330 (Halliday et al., 1988; Marzoli et al., 2000; Rankenburg et al., 2005) and have been well exemplified
331 in the felsic rocks of the Bamenda Mountains associated to the studied mafic lavas (Kamgang et al.,
332 2007).

333
334 As it is evident from the diagrams of the figure 10, when MgO decreases by crystal fractionation
335 process, La/Nb and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios increase, suggesting a contamination process by a high La/Nb -
336 $^{87}\text{Sr}/^{86}\text{Sr}$ component, characteristic of the upper continental crust. The most primitive lavas (MgO > 6
337 wt. %) have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.703 and 0.7034 and La/Nb ratios lower than 0.8 while the most
338 differentiated rocks (MgO < 6%) have $^{87}\text{Sr}/^{86}\text{Sr}$ ratio between 0.7038 and 0.7042 and La/Nb ratio
339 higher than 0.85. Among the more differentiated samples are those with the lowest Pb isotopic ratios,
340 suggesting that the crustal contaminant has low Pb isotopic composition, as already observed along the
341 CVL by Halliday et al. (1988) but with less extreme compositions.

342 An important point is the strong correlation between the $^{87}\text{Sr}/^{86}\text{Sr}$ and the age of the samples (Fig. 11),
343 also observed for the $^{143}\text{Nd}/^{144}\text{Nd}$ and the La/Nb ratios. This correlation indicates that the crustal
344 contamination process became less and less important through time in the Bamenda Mountains.
345 Different mechanisms can account for this time evolution. The country rocks can become more and
346 more refractory with time and can be isolated from the circulating magmas due to the crystallization of
347 the previous magmas in the conduit, or a change in the extensional regime of the whole area can
348 facilitate the circulation of the magmas through the crust, thus preventing interactions of these magmas
349 with the country rocks. Whatever the reason for this time evolution, this implies that the most recent
350 volcanic rocks of the Bamenda province are the most interesting to study the nature and the
351 composition of their mantle source because they are the least contaminated by the overlying crustal
352 rocks.

353 7.2. Fractional crystallization

354 Variations seen on Harker diagrams for most Bamenda mafic rocks (Fig. 5, 6), such as moderate MgO
355 contents ($Mg\# \leq 62$), and Ni (1.54-283 ppm) and Cr (0.70-556 ppm) variations imply the occurrence
356 of some fractional crystallization, as primary magmas in equilibrium with mantle olivine are assumed
357 to have Mg# values of 68–72 and high Ni (300–500 ppm), Cr (300-500 ppm) and Co (50-70ppm)
358 contents (Frey et al., 1978, Jung and Masberg, 1998). Only the basanite sample BA68 have Mg# value
359 of 61 and high Ni and Cr contents (283 and 339 ppm respectively), commensurate with a (near-)
360 primary character.

361 A strong decrease of the Ni and Cr concentrations when the Mg# decreases (Fig. 6a, b) suggests the
362 fractionation of olivine and clinopyroxene. This is confirmed by the decrease of the CaO content of the
363 more differentiated magmas, associated to an increase of the SiO₂ content. However, the constant
364 increase of the Al₂O₃ during the differentiation indicates that crystallization of plagioclase was very
365 limited in these samples.

366 As shown in Figure 4, some basanites contain green-core clinopyroxenes which are evidence for high-
367 pressure fractionation at an early stage of the magma evolution (e.g. Duda and Schmincke, 1985; Jung
368 and Hoernes, 2000; Haase et al., 2004; Jung et al., 2006). This can be related to the absence of
369 plagioclase fractionation as shown by the constant increase in Al₂O₃ in the course of fractional
370 crystallization and could indicate that most of the magma evolution in Bamenda occurs at great depth
371 in the upper mantle or the lower crust.

372 Of course, the large age range span by the Bamenda volcanic rocks implies that we are not dealing
373 with a single differentiation trend, and thus precludes any modelling of the fractional crystallization
374 process.

375 7.3. Mantle source of Bamenda mafic lavas

376 Despite evidences of crustal contamination of some of the mafic magmas before their emplacement at
377 the surface, several observations can be emphasized in order to discuss the nature and composition of
378

379 the mantle source of the magmas in the general framework of the CVL, with a special emphasize to
380 the most recent rocks (< 5Ma) which are not significantly affected by the contamination process.

381 In the Zr-Th diagram (Fig. 12) basanites plot on a distinct correlation line, thus calling for subtle
382 differences in the source composition or in the magma genesis process for these rocks, while no
383 difference in the isotopic composition between the basanites and the other mafic rocks is observed
384 (Fig. 9).

385 An important issue to be considered is whether the garnet is a residual phase in the source or not. The
386 existence of residual garnet results in depletion of HREE relative to LREE because of the strong
387 retention of HREE in garnet, as indicated by high partition coefficients of those elements in garnet
388 (e.g. Green et al., 1989; Jenner et al., 1994). The alkaline affinity of the Bamenda mafic lavas is
389 supported by their enrichments in incompatible elements and high fractionation indexes, $(La/Yb)_n$,
390 which range from 10.62 to 22.80. These high values of the fractionation index correspond to depletion
391 in HREE relative to LREE (otherwise, LREE enrichment) and indicate the presence of garnet as a
392 residual phase in the sources of these rocks. It must also be noted that the $(Tb/Yb)_n$ ratios (1.8-3.0) of
393 the Bamenda mafic lavas are high and in the range of values considered to be characteristics of the
394 presence of residual garnet in the mantle source (Wang et al., 2002). However, no difference of this
395 ratio is observed between the basanite and the other mafic lavas, so their chemical differences
396 evidenced from the Zr-Y-Th content is not to be related to the amount of residual garnet in their
397 respective mantle sources.

398 Most of the studied samples show positive Eu and Sr anomalies without evidence for plagioclase
399 accumulation. This feature is not correlated with the Sr isotopic composition of the rocks, so cannot be
400 related to the crustal contamination process as some of the basanites with the highest Eu anomaly have
401 also low Sr isotopic composition ($^{87}Sr/^{86}Sr < 0.7034$). No correlation exists between these anomalies
402 and the Al or Ca content of the lavas, thus precluding cryptic assimilation of plagioclase-rich
403 cumulates in a magma chamber. The occurrence of such lavas at different places and times along the
404 CVL argue also against this assimilation phenomenon.

405 According to these observations, these chemical anomalies are related to the primary composition of
406 the mafic magmas, and so has something to deal with the composition of the mantle source.

407 Two series of rocks, one with high Sr contents and another one with low Sr concentrations have
408 already been observed in the mafic rocks from the nearby Bambouto volcanic province (Marzoli et al.,
409 2000).

410 The existence of these two rocks series with different trace element composition (one with Eu and Sr
411 positive anomalies, and one without these anomalies) but with similar isotopic ratios implies that two
412 mantle sources with different trace element compositions but similar isotopic ratios have contributed
413 to their genesis. This trace element difference can be attributed to differences in the mineralogy of the
414 source. For the Bambouto volcanic province, the high Sr rock series has been explained by the partial
415 melting of an amphibole-bearing metasomatized mantle source (Marzoli et al., 2000).

416 Such a metasomatized mantle has already been observed in Cameroon and especially in mantle
417 xenoliths from the Nyos volcano where amphibole-bearing spinel lherzolites have been described
418 (Temdjim et al., 2004). The high Sr and Ba contents of amphiboles in these mantle samples, as well as
419 their low Zr and Hf contents (Fig. 13a) make them potential candidates for the source of the high-Sr
420 mafic magmas. However, these amphiboles do not show the high Eu content observed in the Bamenda
421 mafic rocks and their participation to the genesis of these magmas do not explain the positive Eu
422 anomalies. The pyroclastic deposits from the Youkou volcano in the Adamawa volcanic field contain
423 mantle xenoliths and especially garnet-bearing pyroxenites. Whole rock chemical composition of these
424 samples (France et al., submitted) shows that they are enriched in Sr and Ba and also possess a
425 positive Eu anomaly in their REE patterns (Fig. 13b), as well as low Zr and Hf contents. Furthermore,
426 the Sr and Nd isotopic composition of the pyroxenite xenoliths is not very different from that
427 measured in the Bamenda mafic volcanic rocks. Partial melting of these pyroxenites and mixing of
428 these magmas with magmas generated in associated garnet lherzolites can explain the chemical and
429 isotopic characteristics of the Bamenda volcanic rocks. These pyroxenites also contain apatite, which
430 is in good agreement with the observation that the Bamenda basalts with the highest Eu and Sr
431 anomalies also have the highest P₂O₅ contents. The similarity of the isotopic compositions of the two

432 different kinds of mafic rocks in the Bamenda Mountains is consistent with a recent origin for the
433 mantle source enriched in Eu, Sr and Ba or a time evolution of the source with low Rb/Sr and
434 moderate Sm/Nd ratios, which is the case for the Youkou pyroxenites.

435 In the different isotopic diagrams of the figure 9, the Bamenda mafic rocks form a linear array
436 stretching from a low Sr and high Nd and Pb isotopic ratio component close to the HIMU mantle
437 composition to a high Sr and low Nd-Pb isotopic ratio component. This last component probably
438 represents the composition of the local continental crust and highlights the crustal contamination
439 processes which have affected some of these samples. On the other hand, the high Pb isotopic
440 composition component has already been discussed and is commonly related to the enrichment of the
441 subcontinental lithospheric mantle by metasomatic fluids originated from the St Helena mantle plume,
442 or to the direct incorporation of the St Helena plume material into the lithospheric mantle. All the
443 geochemical data acquired on the CVL are consistent with an origin of the magmas from the mixing
444 between asthenospheric melts and HIMU-type melts formed by partial melting of the lithospheric
445 mantle (Halliday et al., 1988; Marzoli et al., 2000; Rankenburg et al., 2005; Yokoyama et al., 2007).

446 Our data are consistent with the involvement of such an HIMU mantle component in the genesis of the
447 Bamenda mafic rocks all along the 25 Ma of duration of the volcanism in this province. Furthermore,
448 the presence of the Sr and Eu anomalies in most of the Bamenda rocks allows us to constrain the
449 nature of the St Helena plume fingerprint in the mantle. As shown before, these trace element
450 anomalies have been observed in garnet pyroxenites from the Youkou volcano in the CVL. France et
451 al. (submitted) have interpreted these pyroxenites as the crystallization at mantle depth of magmas
452 formed by melting of the St Helena mantle plume. From our data, we can argue that the high Sr-Eu
453 volcanic rocks of the CVL are probably formed by the interaction of asthenospheric melts with melts
454 produced by the partial melting of garnet pyroxenites resulting from the crystallization of mafic
455 magmas formed at the time when the St Helena was below this area of the CVL. Magmas without
456 these Sr and Eu anomalies but with similar isotopic composition are probably the result of interaction
457 between asthenospheric melts and melts formed by the melting of a metazomatized lithospheric mantle
458 as already noticed in the other volcanic provinces of the CVL. The Bamenda mafic rocks allow for the

459 first time to evidence the participation of old pyroxenitic veins in the genesis of magmas along the
460 CVL.

461

462 8. Conclusion

463 - Mafic magmas from the Bamenda Mountains range from basanites and hawaiites to basalts and
464 mugearites with an alkaline affinity.

465 - They were emplaced on a large time range from 18 Ma to very recent time but despite this large time
466 range, there are no important chemical differences between the oldest and the youngest mafic rocks.

467 - Some samples, especially the oldest ones, show some evidences of contamination during interaction
468 with the continental crust.

469 - As a whole, the Bamenda mafic rocks were formed in the garnet stability field and resemble OIB
470 generated above a mantle plume, but some rocks show Eu and Sr positive anomalies not related to
471 plagioclase accumulation. These peculiar chemical features are consistent with the partial melting of
472 garnet pyroxenites similar to that found as xenoliths in the Adamawa region and which possess the
473 same chemical characteristics as well as similar isotopic composition.

474 - We propose that the magmas with Eu and Sr anomalies are generated by interactions of
475 asthenospheric melts with melts formed by the melting of such garnet pyroxenites, while the other
476 magmas are formed by the partial melting of metasomatized lherzolites. The pyroxenites probably
477 represent melts crystallized at mantle depth when the St Helena mantle plume was active in this
478 region.

479

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725 **Figure captions**

726

727 Fig. 1. (a) Location of Cameroon in Africa; (b) location of the Bamenda Mountains along the
728 Cameroon line showing the location (black) of the Cameroon line volcanic rocks (adapted after
729 Halliday et al. (1988)); the boundary between continental and oceanic crust is taken from Emery and
730 Uchupi (1984) and oceanic transform faults are from Sibuet and Mascle (1978); (c) Simplified
731 geological map of the Bamenda area. C = Cameroon; COA = West African craton; CC = Congo
732 craton; CT = Tanzanian craton; CK = Kalahari craton. African cratons are after Kampunzu and Popoff
733 (1991).

734

735 Fig. 2. Total alkali vs. silica (TAS) classification diagram for Bamenda mafic lavas (Le Bas et
736 al., 1986). Inset: Na₂O vs. K₂O subdivision diagram of the alkaline series (Middlemost, 1975). The
737 alkaline – subalkaline dividing line is from Miyashiro (1978).

738

739 Fig. 3. a) End-member compositional variation in clinopyroxene (classification after Morimoto et al.,
740 1988) and olivine from Bamenda mafic lavas. b) An-Ab-Or triangular plot (Smith and Brown, 1988)
741 showing the composition of plagioclase from Bamenda mafic lavas. Symbols are as follows: triangles
742 = basanite; rectangles = basalt; circles = hawaiite; inverted triangles = mugearite.

743

744 Fig. 4. Major-element variations (in wt. %) along rim to core traverses across one green-core
745 clinopyroxene phenocryst (a) (sample BA37), two typical diopsides (b) and (c) (samples BA37 and
746 BA61), and one plagioclase phenocryst (d) (Sample BA26).

747

748 Fig. 5. Variation of major element oxides (in wt. %) in the Bamenda mafic lavas as a function of Mg#.

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750 Fig. 6. Variation of selected trace elements (in ppm) versus SiO₂ (in wt. %) or Mg# for Bamenda mafic
751 lavas.

752

753 Fig. 7. Chondrite normalized rare earth element patterns of Bamenda mafic lavas a): basanite, b):
754 basalt, c): hawaiiite, d): mugearite. Values of chondrite are from McDonough (2003). Also plotted are
755 patterns of OIB, N-MORB and E-MORB from Sun and McDonough (1989).

756

757 Fig. 8. Primitive mantle-normalized multi-element patterns for the Bamenda mafic lavas. a): basanite,
758 b): basalt, c): hawaiiite, d): mugearite. Patterns for OIB, E-MORB and N-MORB are given for
759 comparison. The normalization values are from McDonough (2003) and values of OIB, N-MORB and
760 E-MORB from Sun and McDonough (1989).

761

762 Fig. 9. ⁸⁷Sr/⁸⁶Sr vs. ¹⁴³Nd/¹⁴⁴Nd (a); ²⁰⁶Pb/²⁰⁴Pb vs. ⁸⁷Sr/⁸⁶Sr (b); ¹⁴³Nd/¹⁴⁴Nd vs. ²⁰⁶Pb/²⁰⁴Pb (c);
763 ²⁰⁶Pb/²⁰⁴Pb vs. ²⁰⁷Pb/²⁰⁴Pb (d) diagrams for Bamenda mafic rocks. The fields of HIMU, OIB, FOZO,
764 Dupal OIB, EMI and EM2 are from Hamelin and Allègre (1985), Zindler and Hart (1986), Hart (1984,
765 1988), Hart et al. (1992), Hawkesworth et al. (1984), Weaver (1991). Low Nd (LoNd) array and
766 northern Hemisphere Reference Line (NHRL) are from Hart (1984). The fields for Atlantic-Pacific
767 MORB are from Hamelin and Allègre (1985) and references therein. Literature data for CVL are
768 compiled from Halliday et al. (1988, 1990), Lee et al. (1994), Marzoli et al. (1999, 2000), Rankenburg
769 et al. (2005), Yokoyama et al. (2007); Nkouandou et al. (2008) Nkouathio et al. (2008).

770

771 Fig. 10. ⁸⁷Sr/⁸⁶Sr and La/Nb variations vs. MgO contents of the Bamenda mafic rocks.

772

773 Fig. 11. Present-day ⁸⁷Sr/⁸⁶Sr vs. age of the mafic samples from Bamenda. Symbols as in figure 10.

774

775 Fig. 12. Zr vs. Th for mafic lavas of Bamenda Mountains.

776

777 Fig. 13. Trace element composition of some Bamenda basalts with Sr and Eu positive anomalies
778 compared to the composition of pargasites from amphibole-bearing spinel lherzolites from the Nyos
779 locality (Temdjim et al., 2004) and pyroxenites from Youkou (France et al., submitted).

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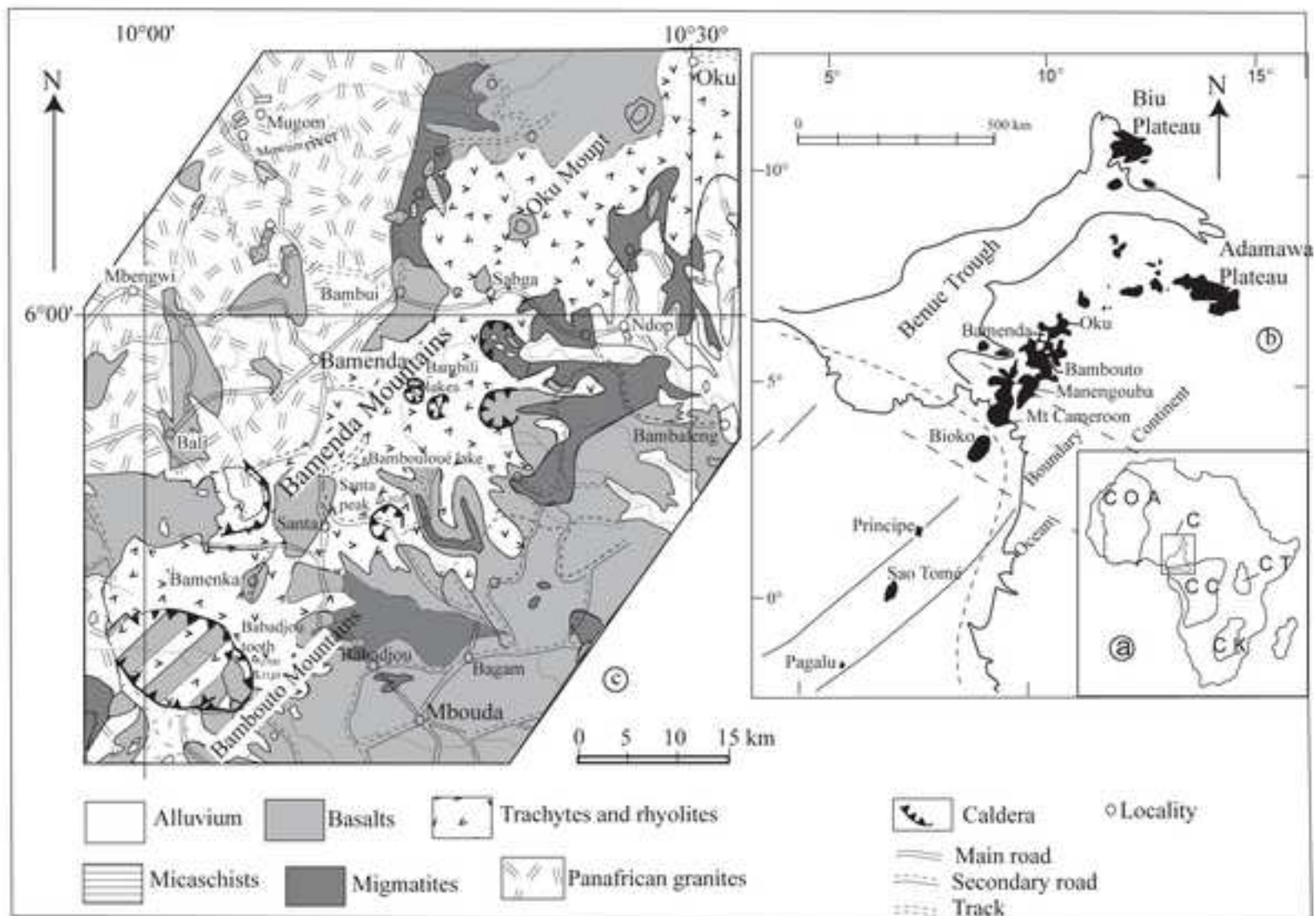


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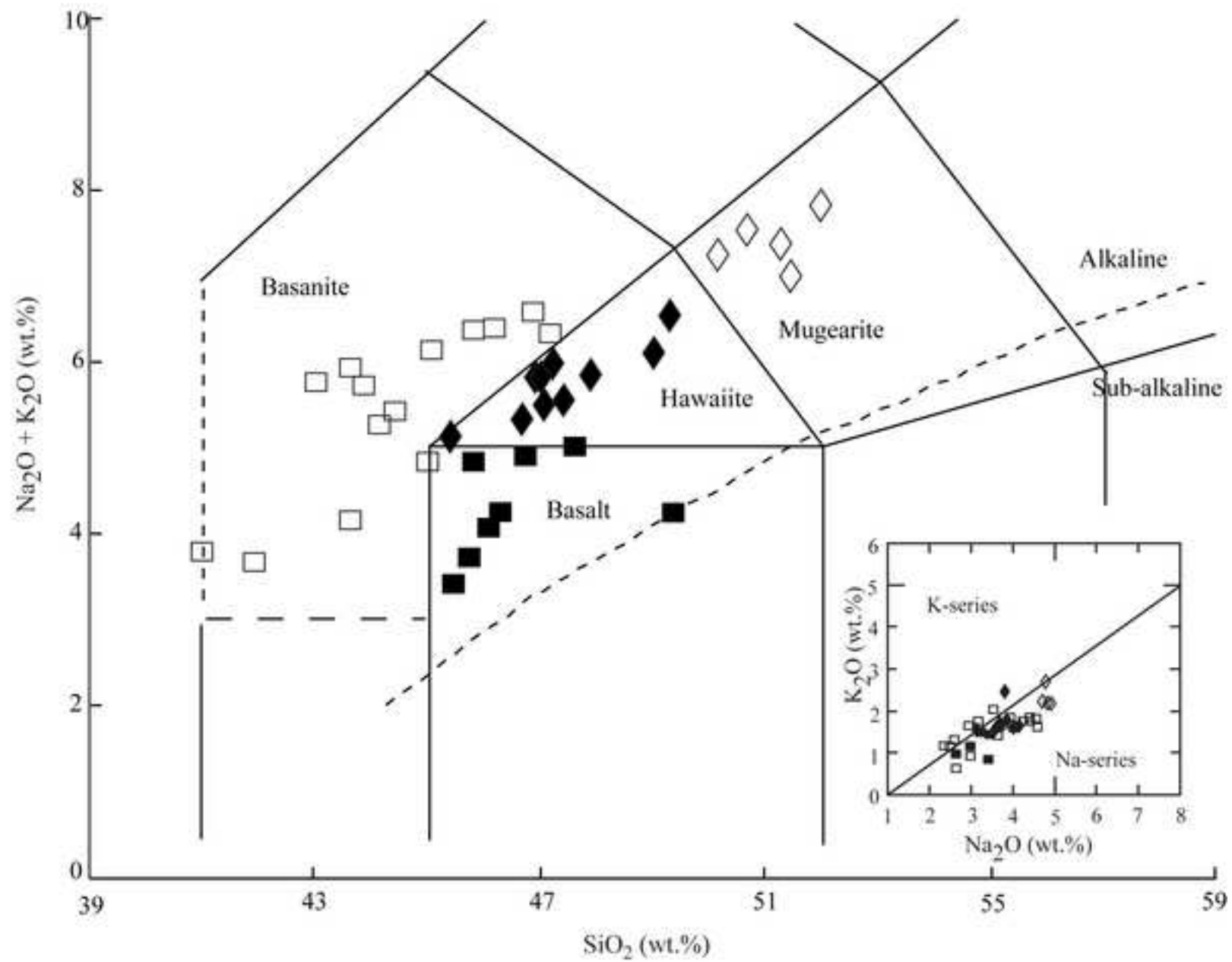


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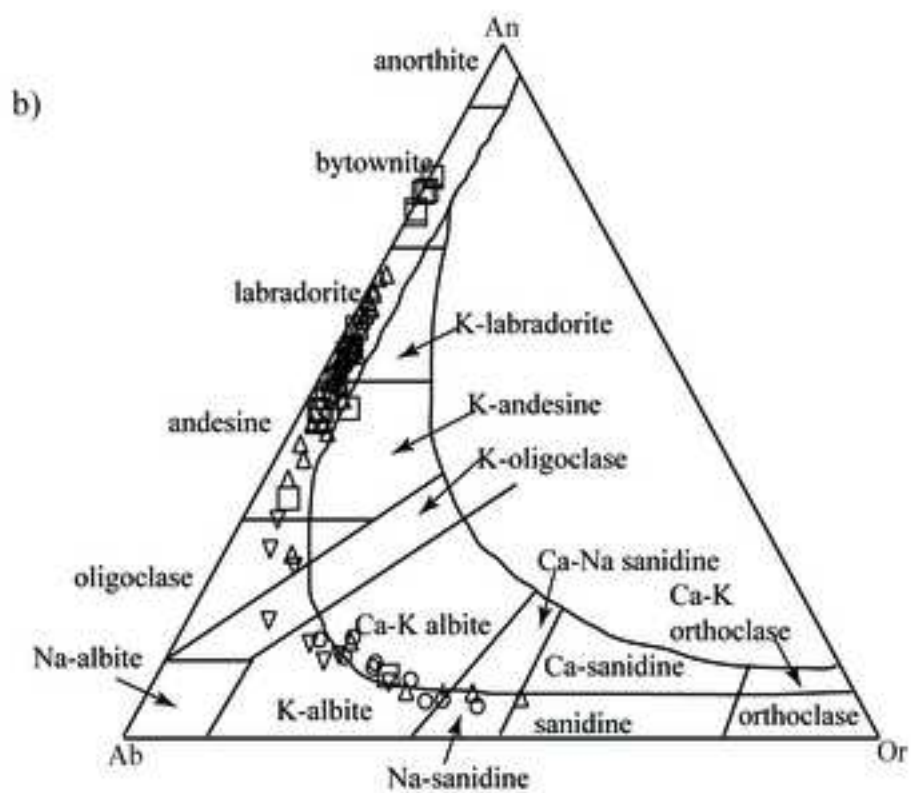
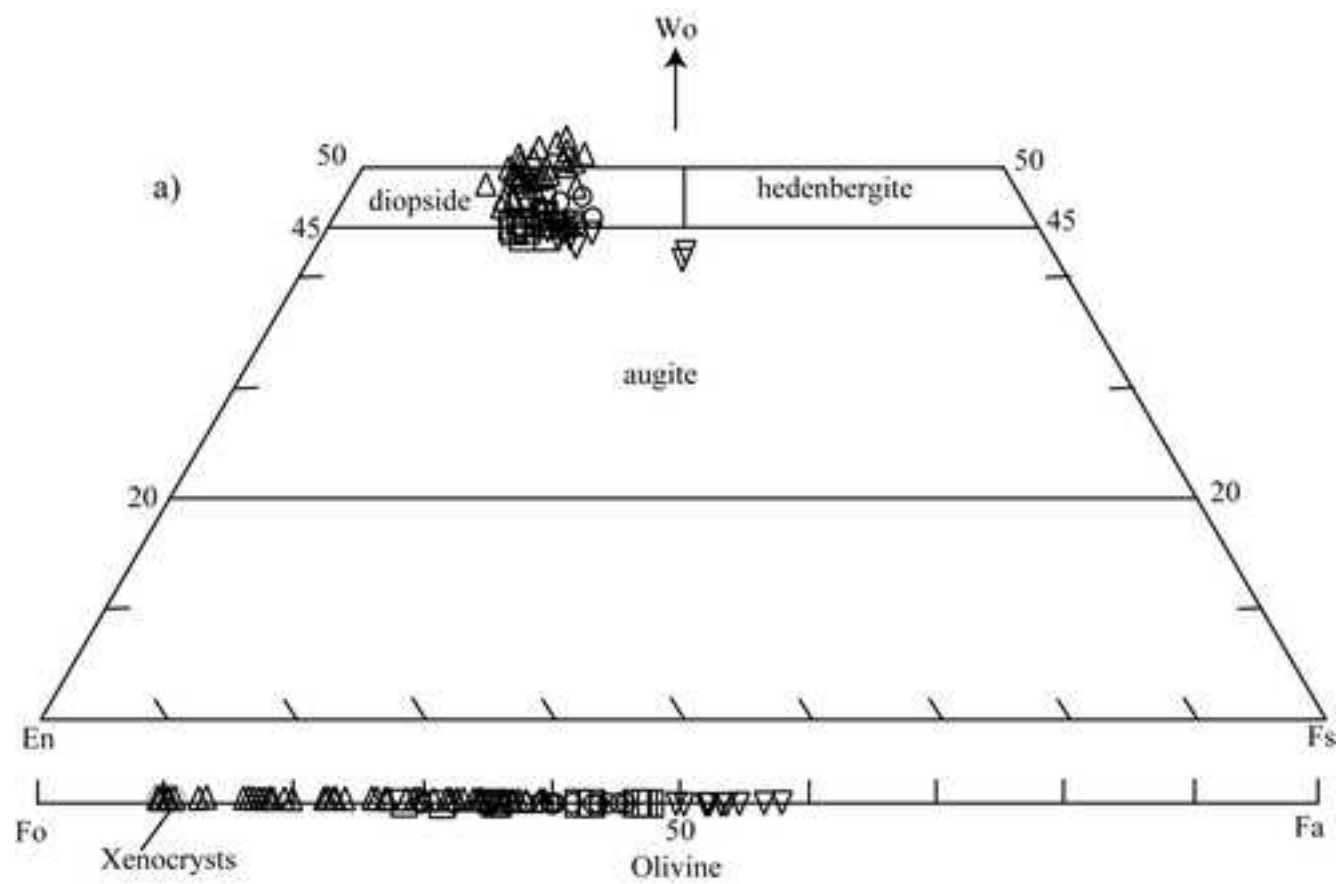


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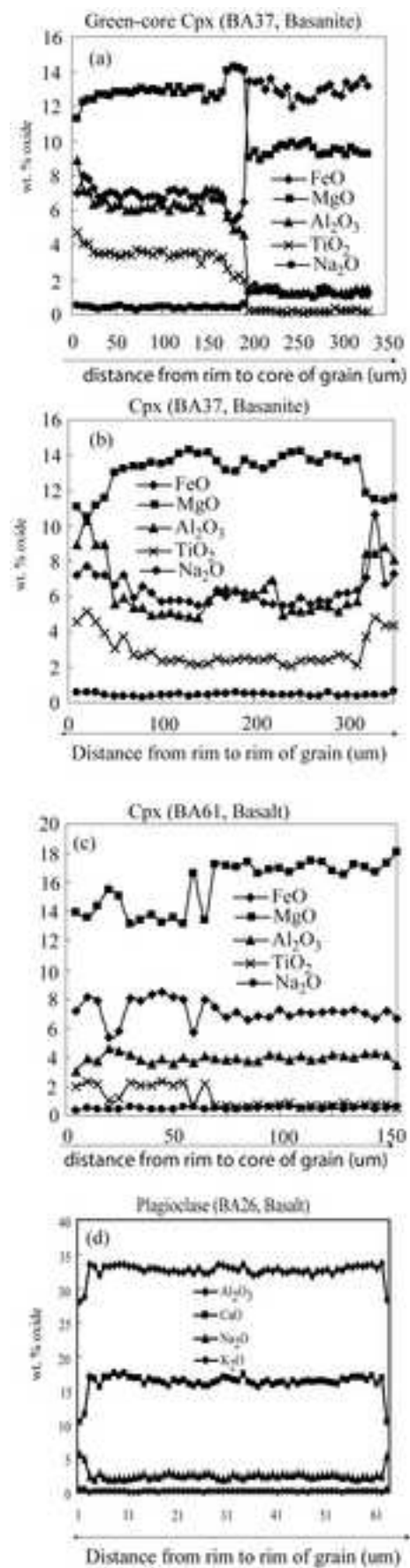


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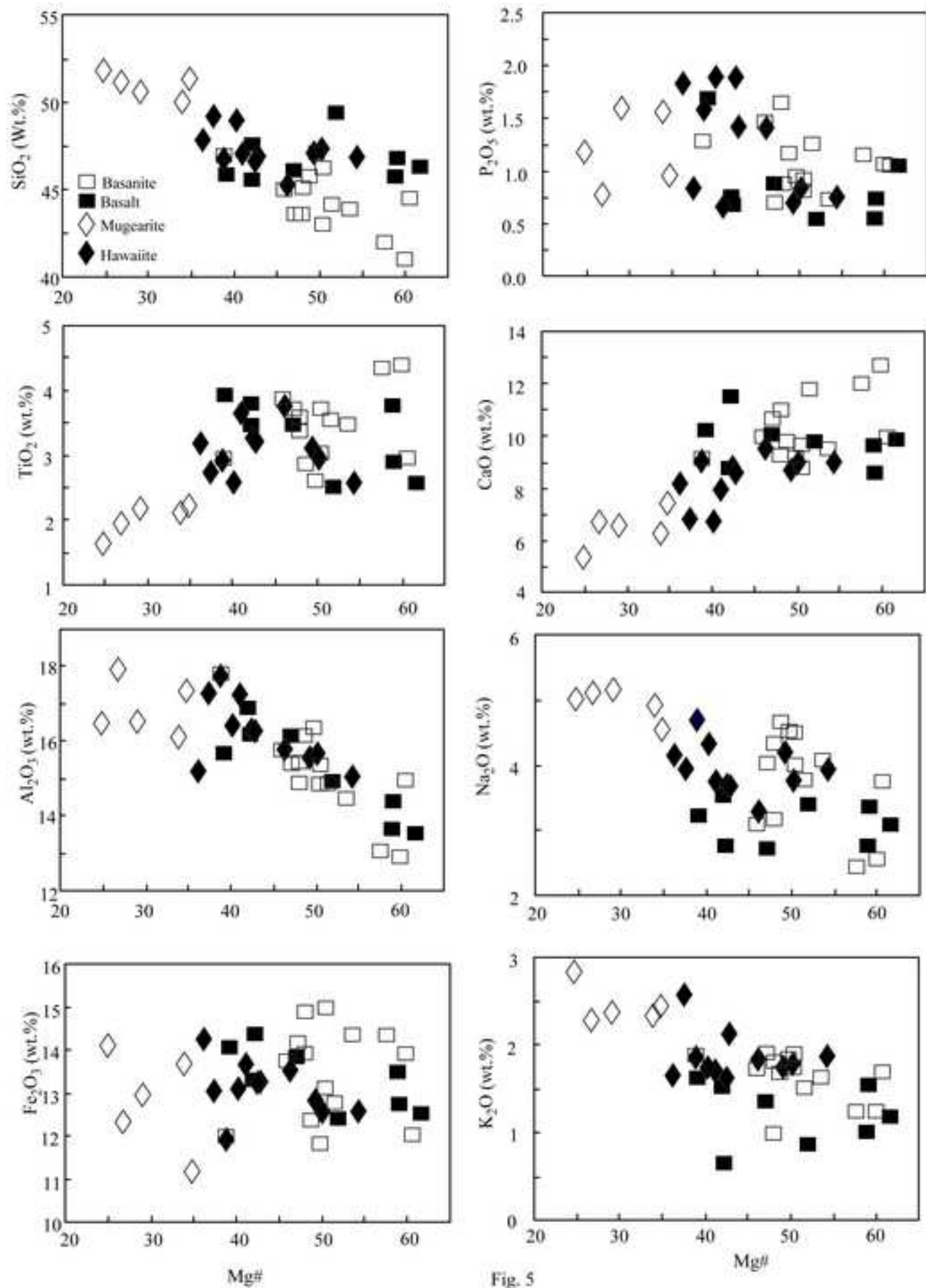


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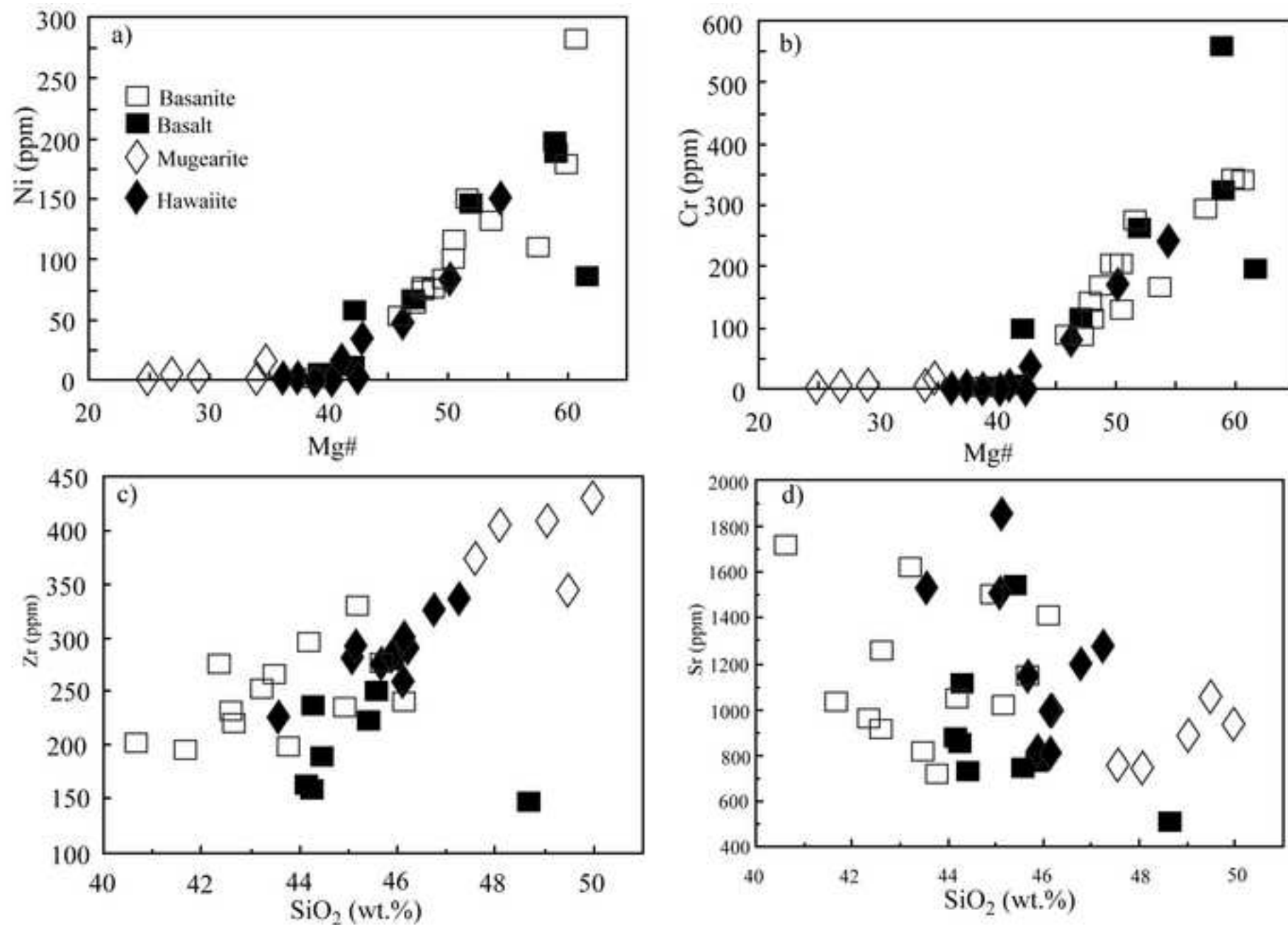


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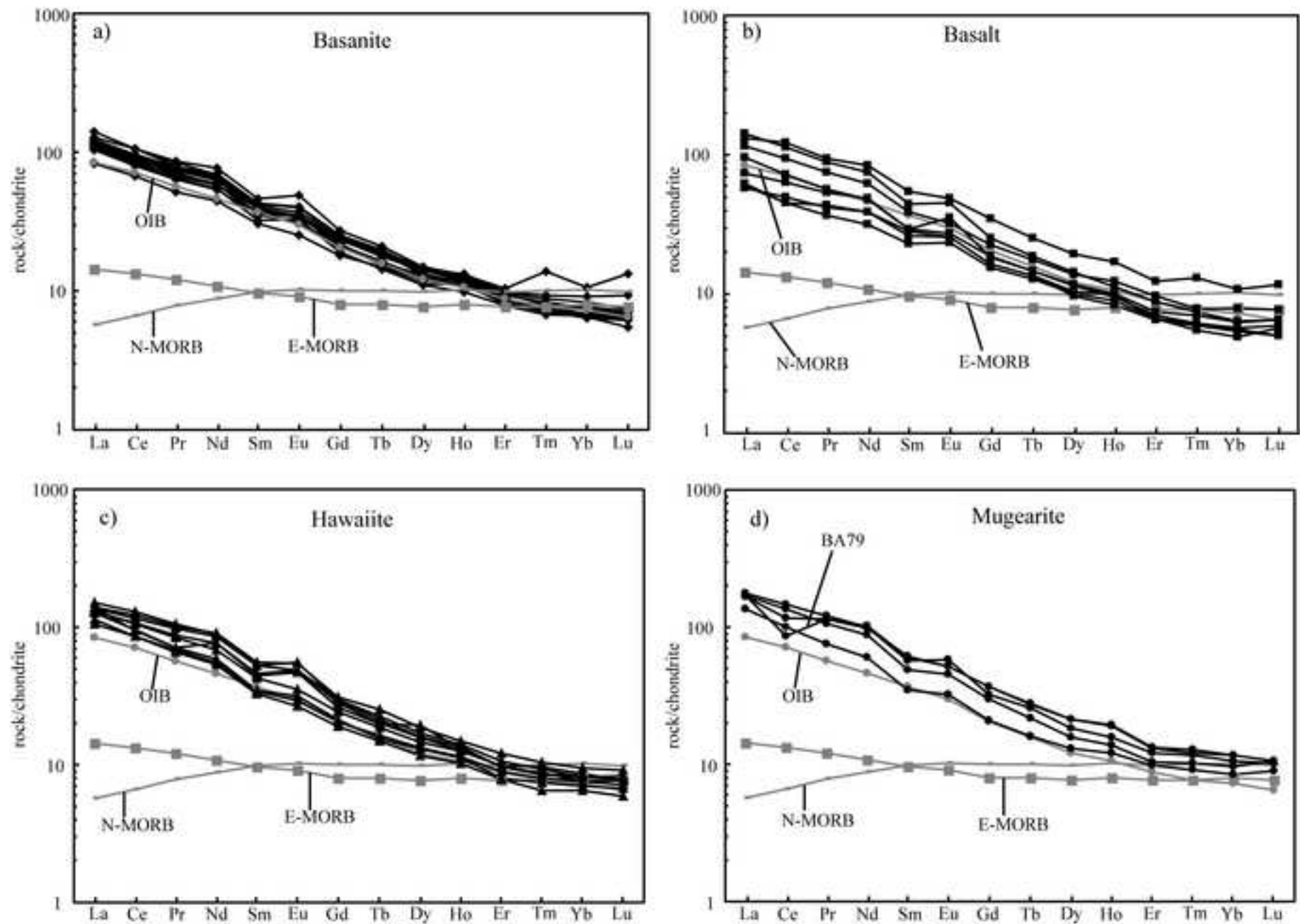


Fig. 7

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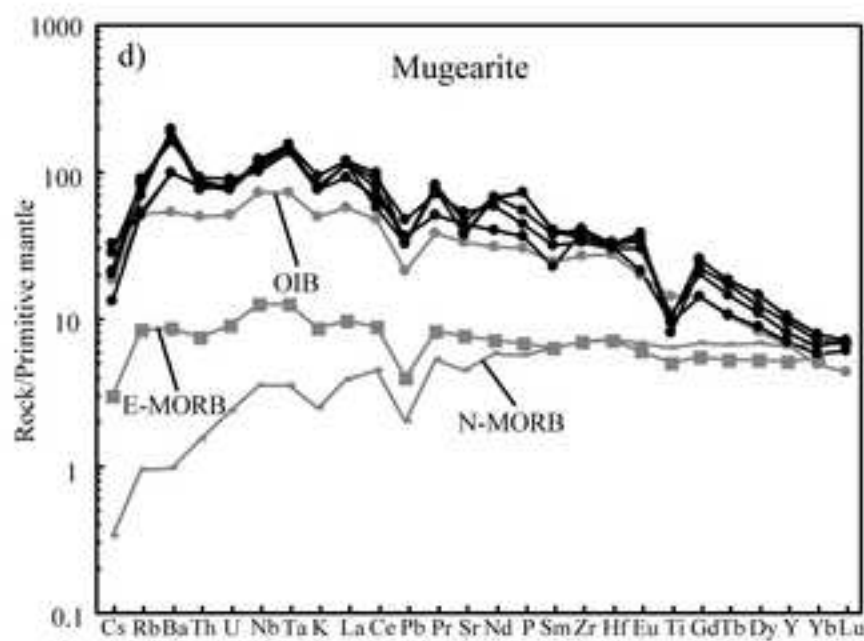
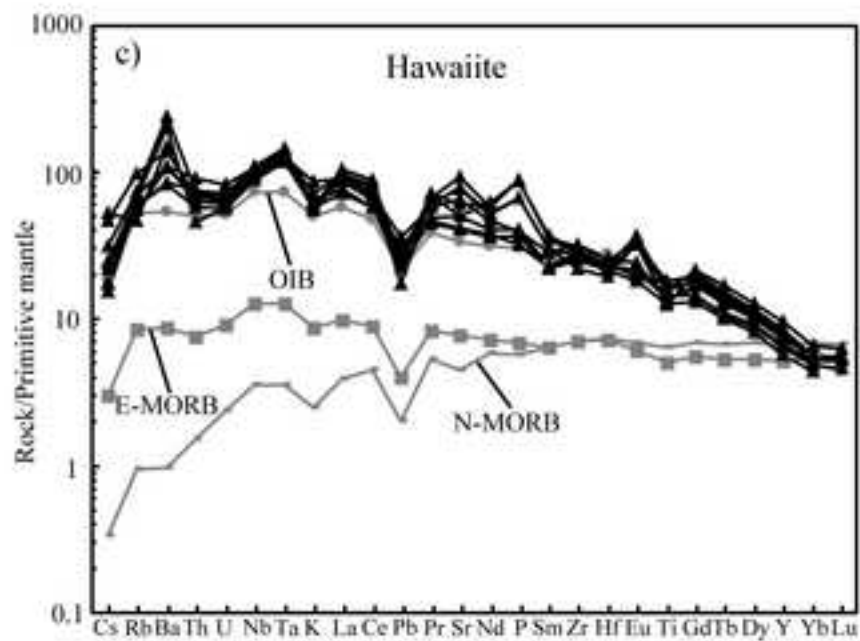
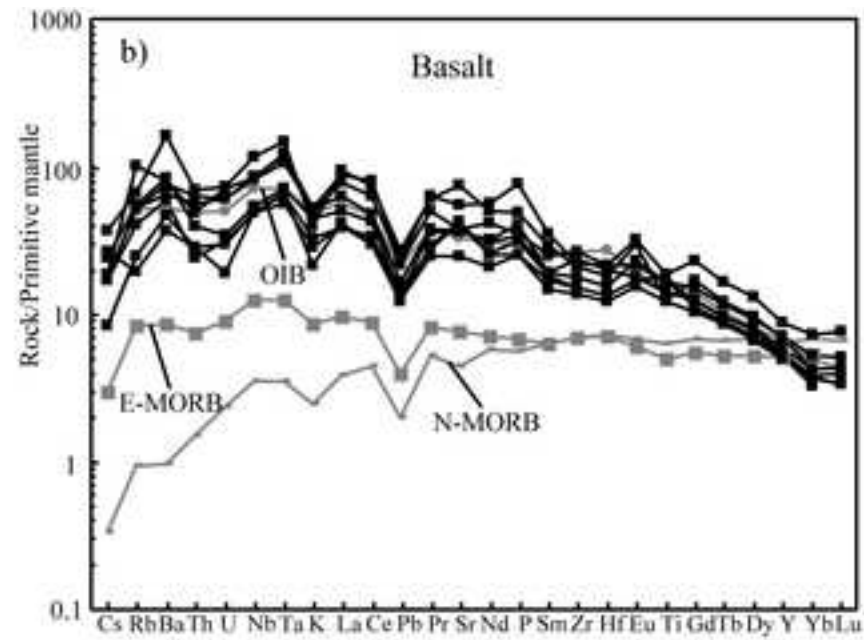
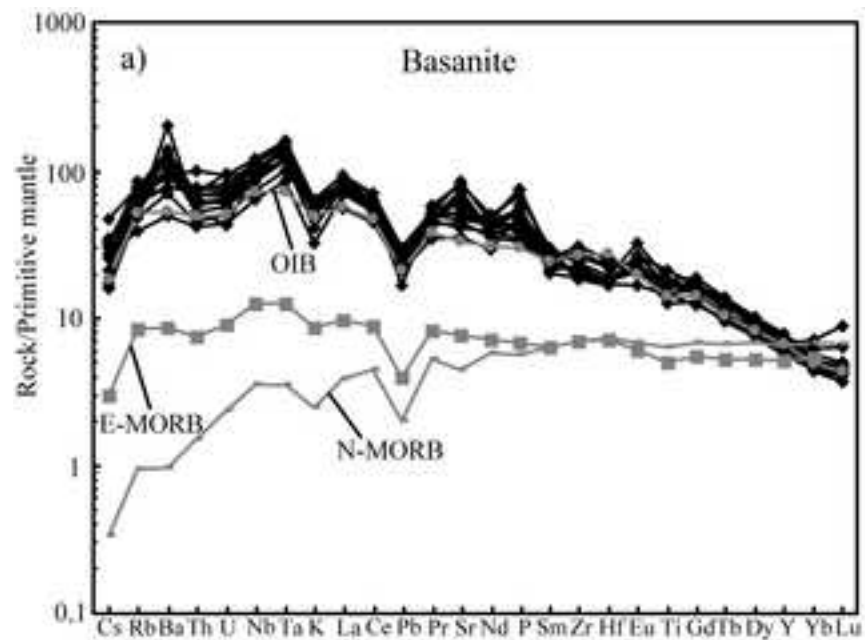


Fig. 8

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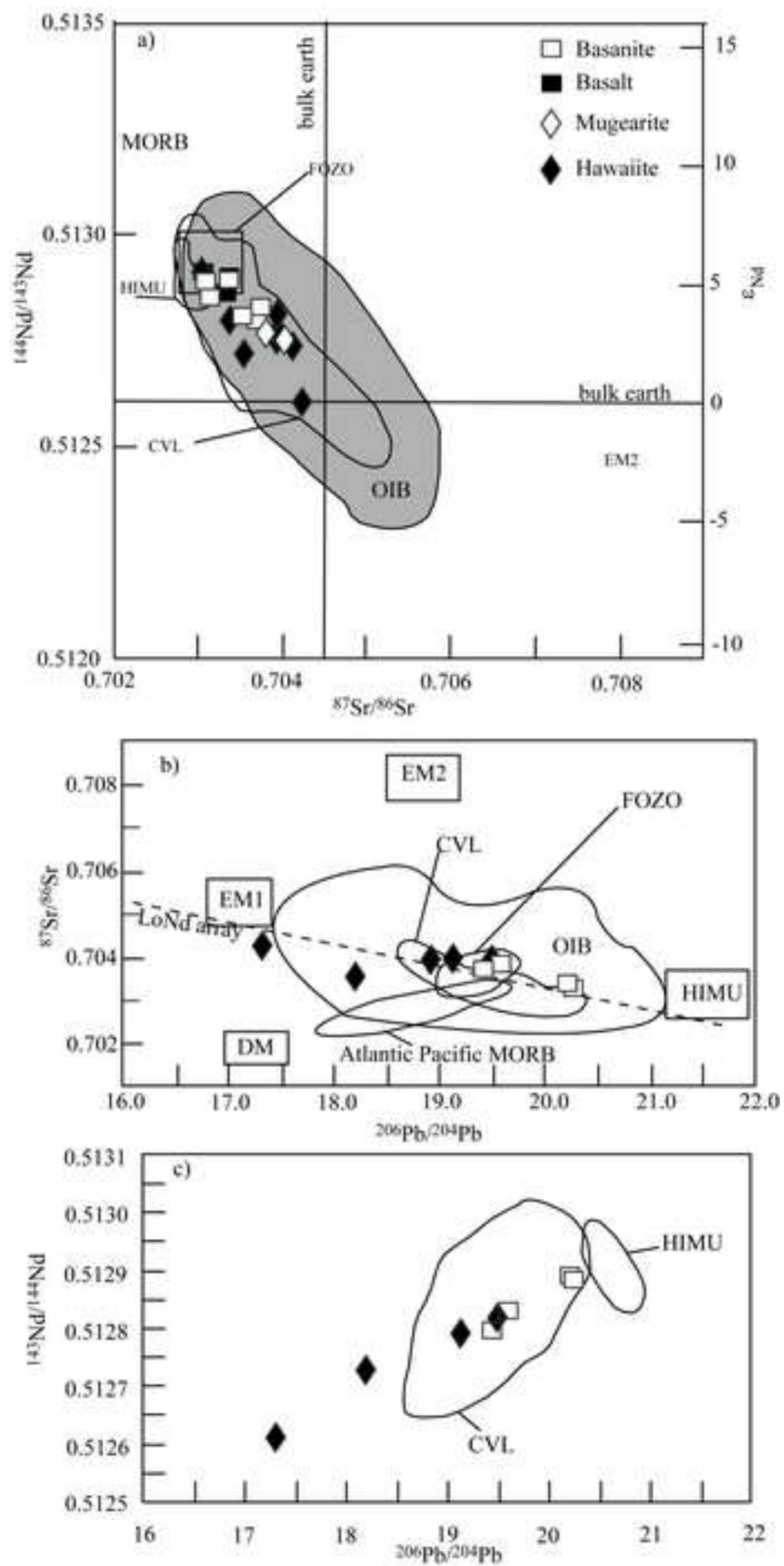


Fig. 9

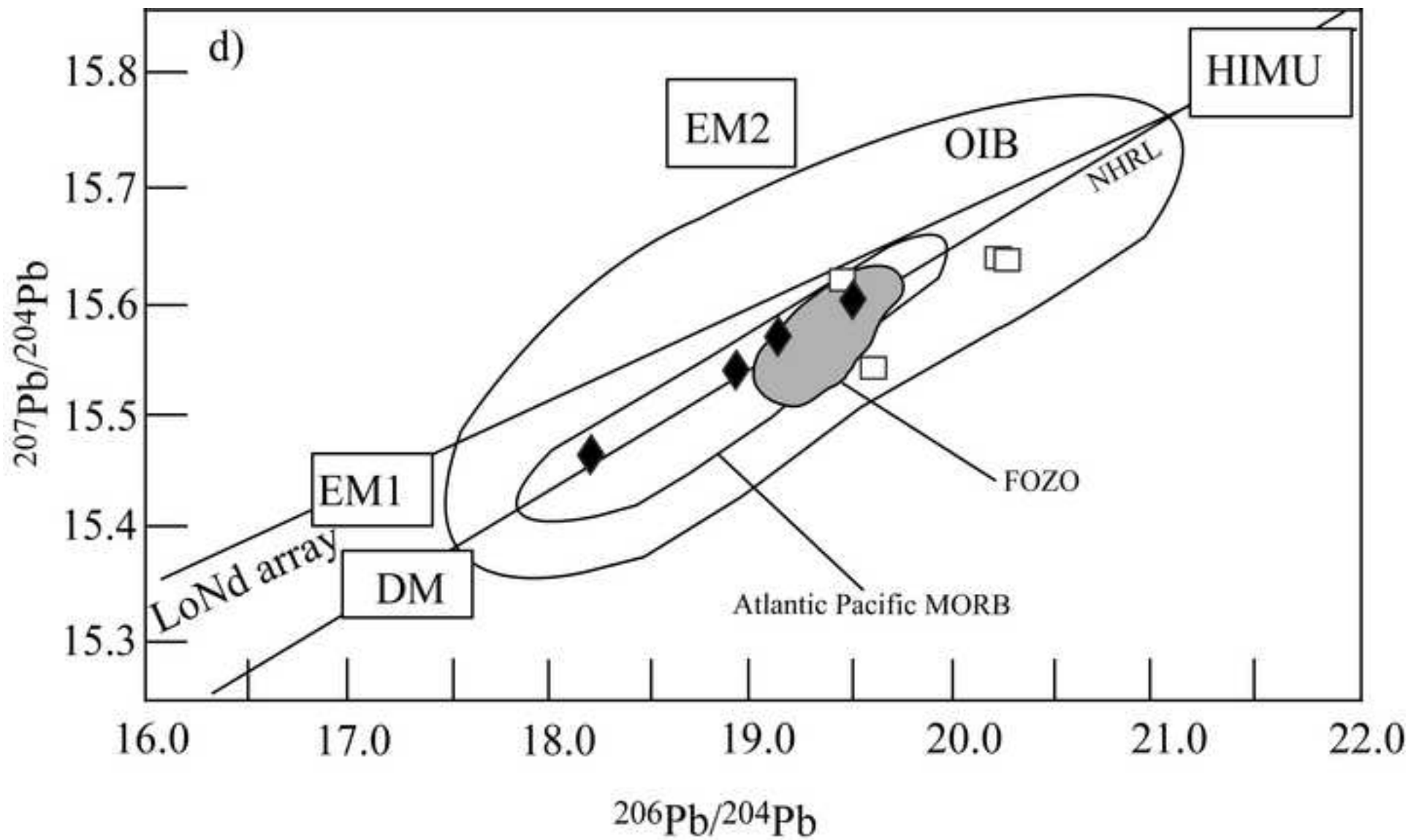


Fig. 9 (continued)

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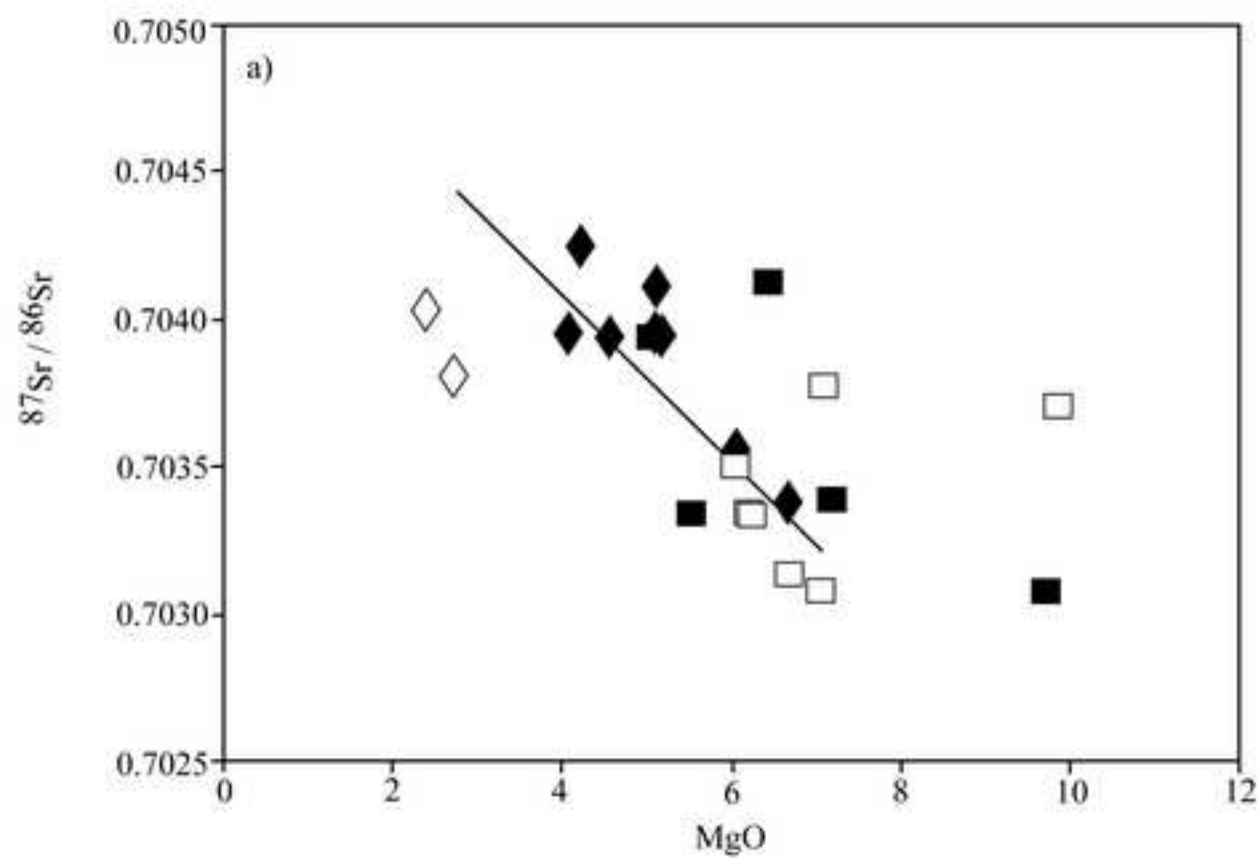
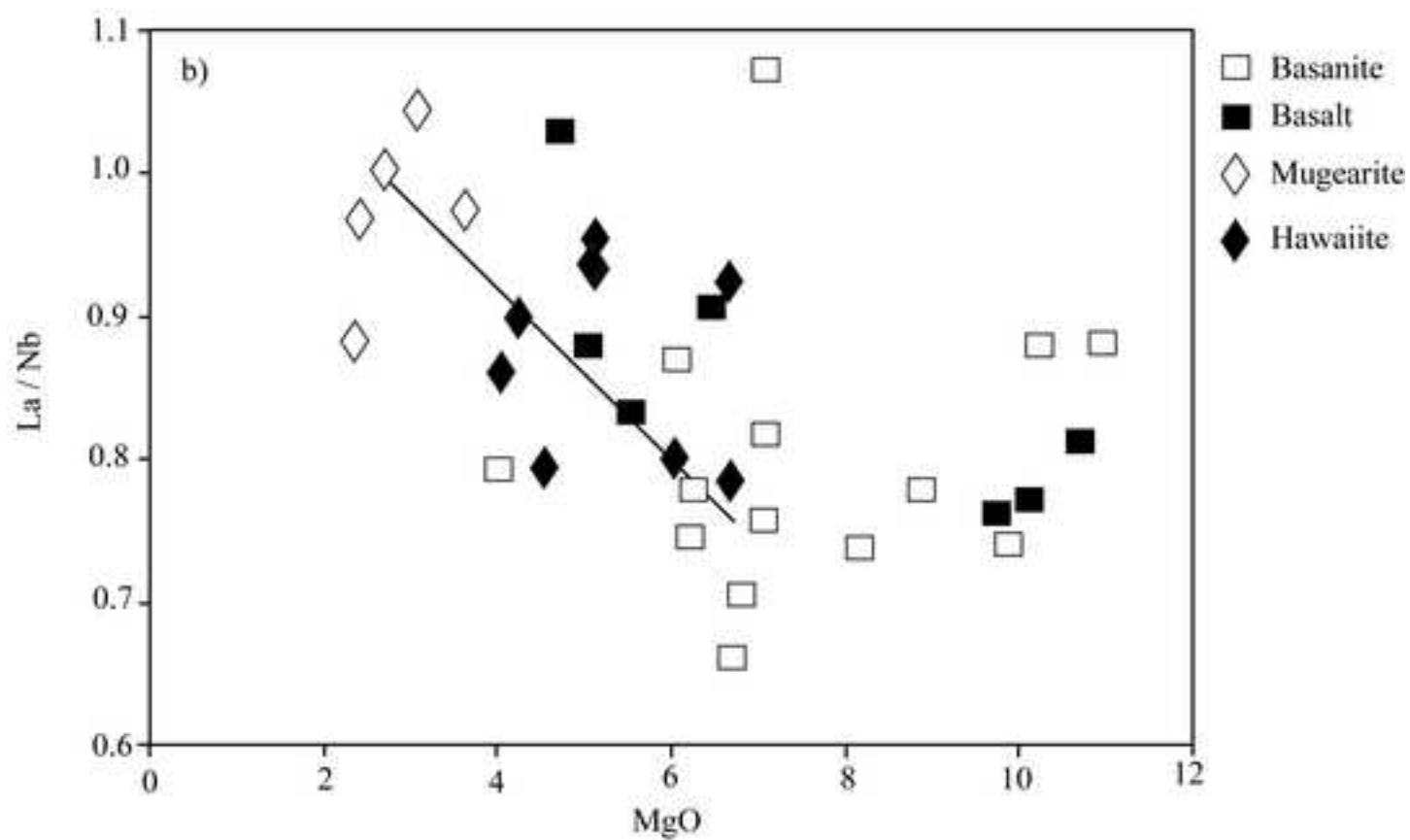


Fig. 10

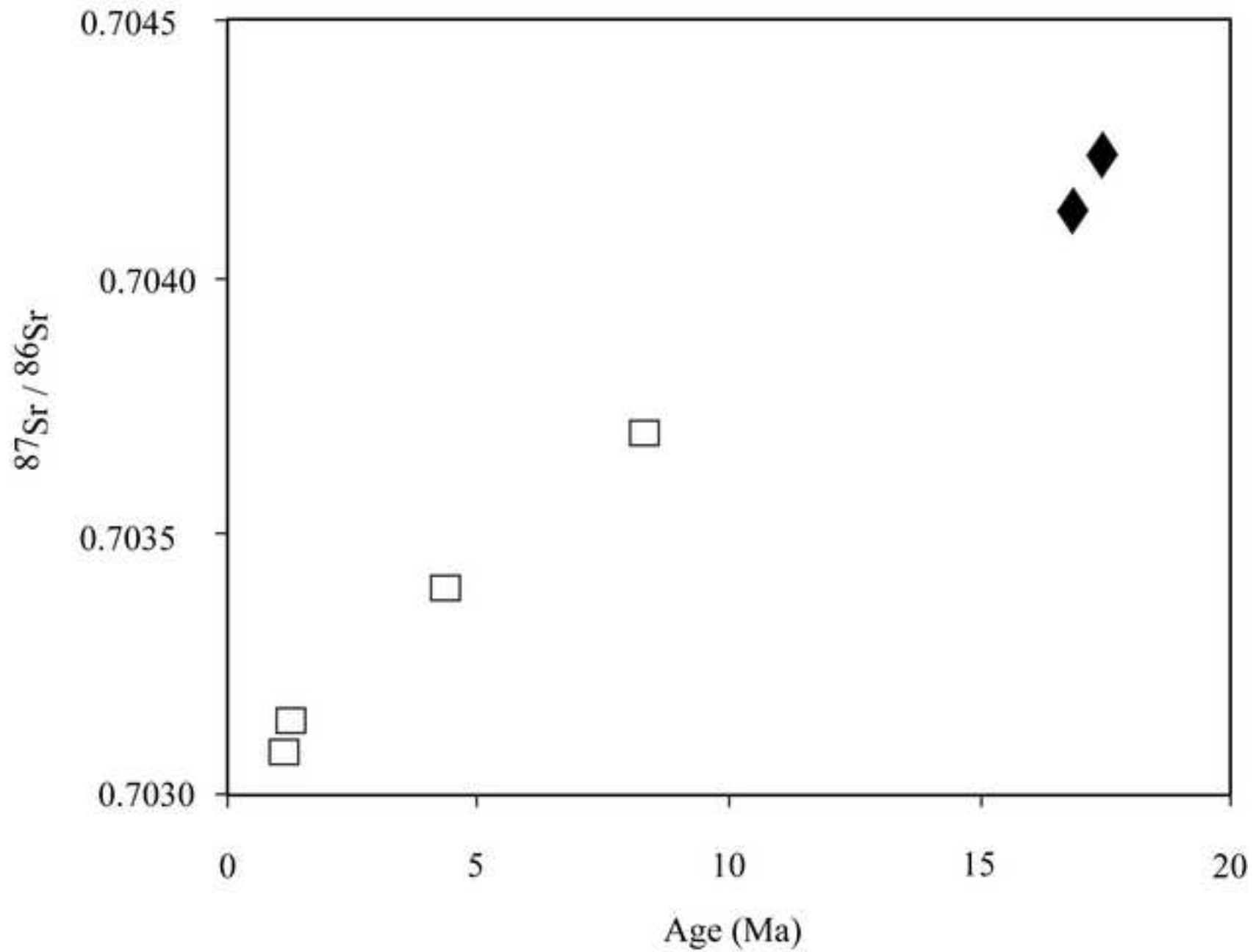


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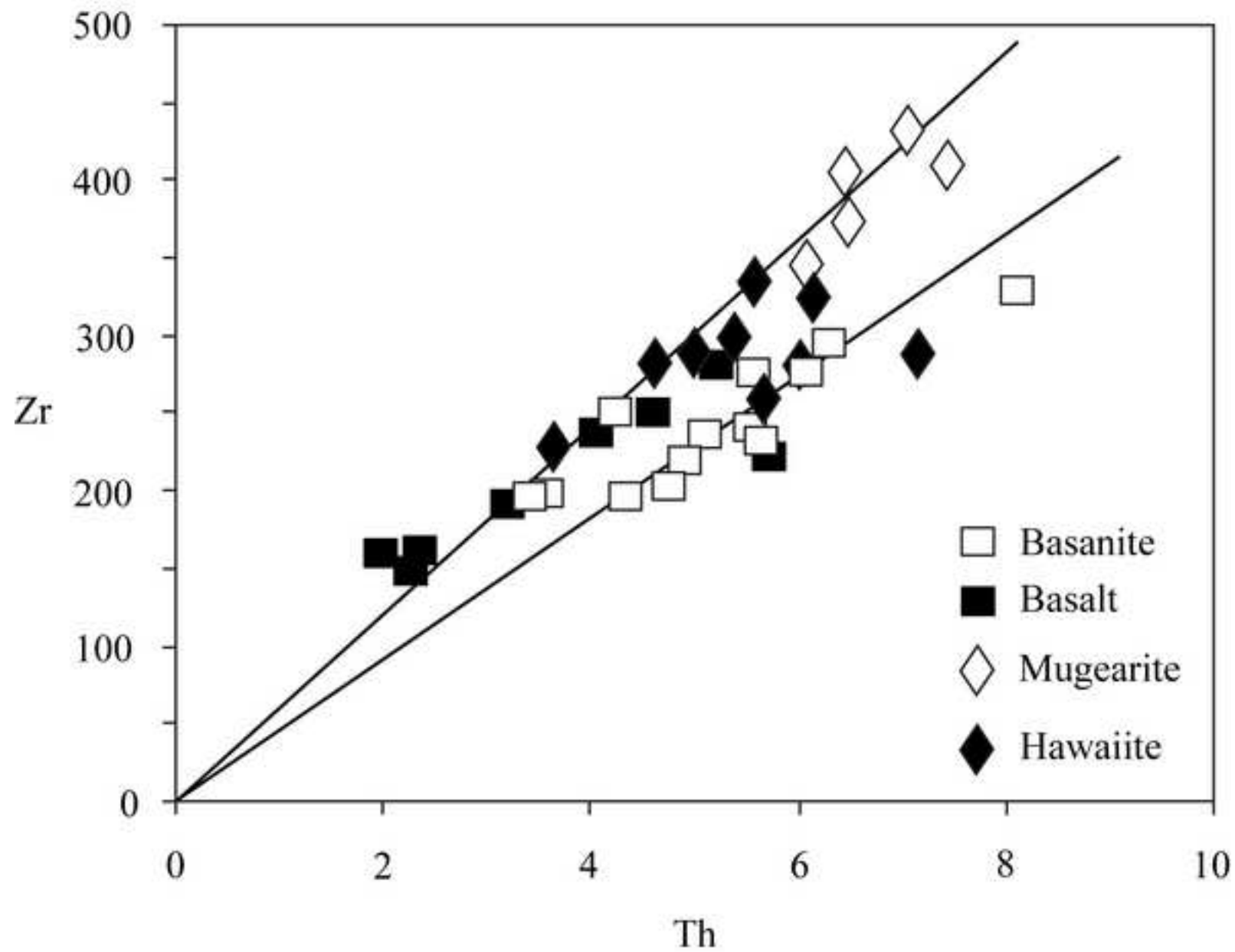


Fig. 12

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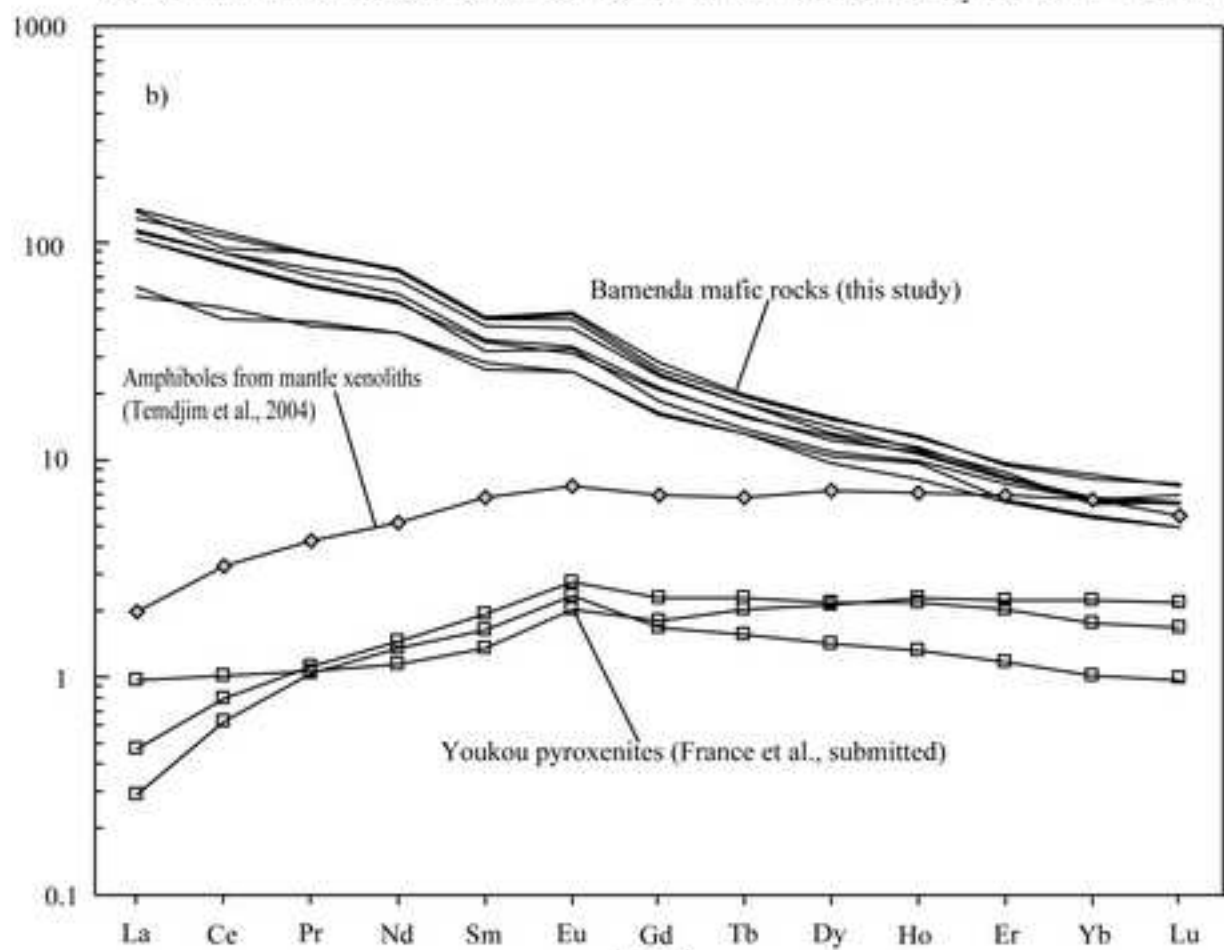
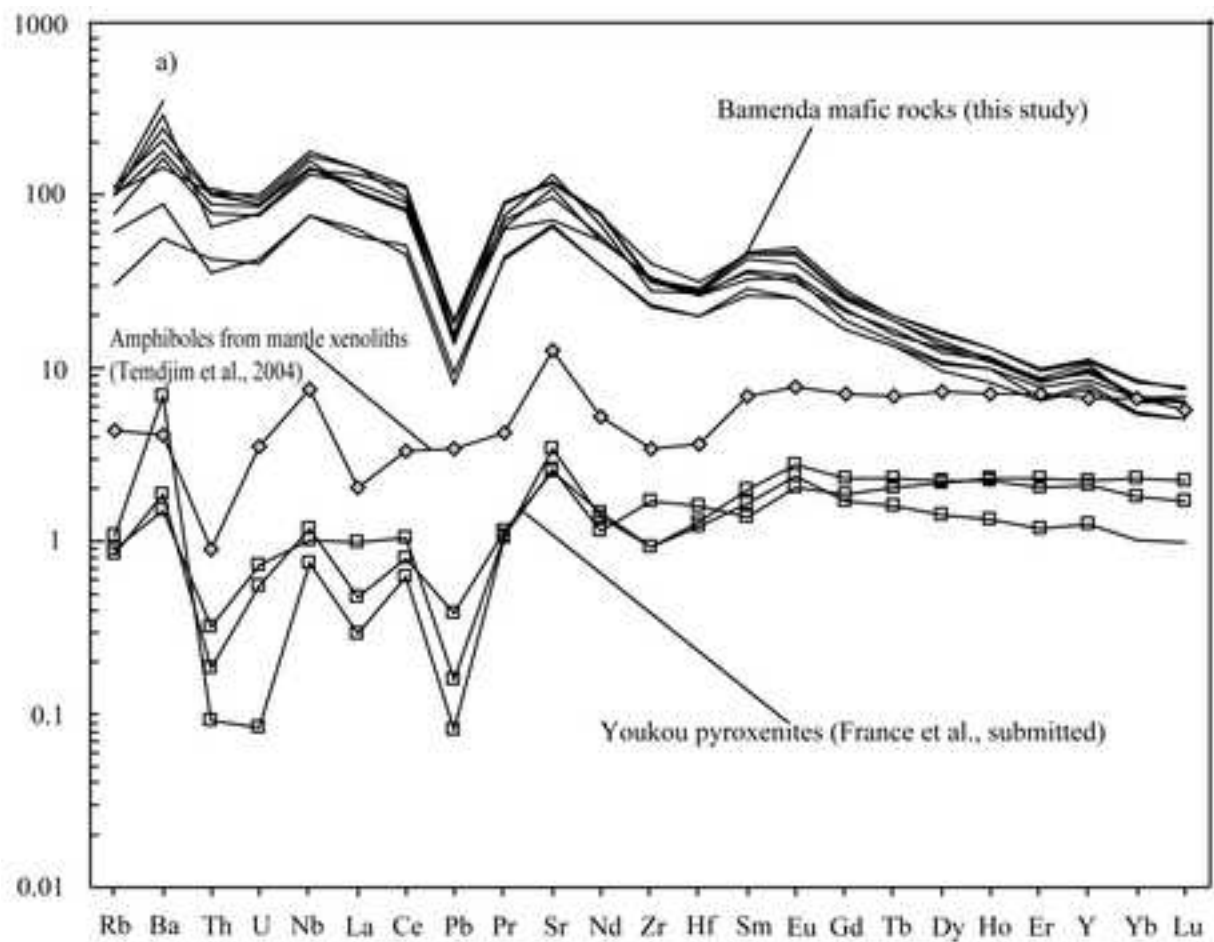


Fig 13

Table1. Main mineralogical features for the different rock types.

	Basanite	Basalt	Hawaiite	Mugearite
Olivine (Fo)	62-87	52-85	54-70	42-50
Ca-rich Cpx				
Wo	46-52	44-46	44-47	41-45
En	34-41	38-41	34-39	29-38
Fs	11-15	15-16	14-17	17-29
Mica (X_{Mg})	-	-	Phlogopite 0.70-0.82	-
Feldspathoids	Nepheline	-	-	-
Plagioclase (An)	1-66	9-81	5-61	8-55
Titanomagnetite (Usp)	48-86	77-85	76-79	73-77
Apatite			+	+
Calcite		+	+	+

Table 2. Major and trace element data for representative samples of the Bamenda mafic lavas.
Oxides as wt.%; trace elements in ppm.

Basanite	BA04	BA31	BA43	BA68	BA81
wt. %					
SiO ₂	44.17	43.47	45.65	43.75	42.36
TiO ₂	3.32	3.45	2.86	2.91	3.66
Al ₂ O ₃	15.14	14.33	17.31	14.71	14.64
Fe ₂ O ₃	13.64	14.21	11.67	11.86	14.78
MnO	0.2	0.2	0.2	0.2	0.17
MgO	6.8	8.86	4.01	9.85	8.14
CaO	9.05	9.38	8.88	9.83	9.54
Na ₂ O	4.25	4.04	4.59	3.69	3.95
K ₂ O	1.77	1.62	1.82	1.66	1.72
P ₂ O ₅	0.86	0.72	1.55	1.04	0.91
LOI	0.41	-0.2	1.04	0.33	0.2
Total	99.61	100.08	99.58	99.83	100.07
Mg #	48	54	39	61	50
ppm					
Nb	76.5	63.5	77.8	49.6	70.4
Zr	296	266	277	199	276
Y	32.5	28.7	32.5	26.6	29.4
Sr	1050	819	1154	718	958
Rb	46.5	39.6	42.8	28.9	37.0
Ga	25.1	21.9	6.5	22.4	21.5
Ni	76.6	131	2.2	283	100
Co	40	46.6	19.6	56	46.9
Cr	115	167	1.6	339	131
Sc	20.2	22.8	13.9	1.0	17.6
V	227	230	134	219	248
Cs	0.99	0.7	0.5	0.5	0.4
Ba	599	511	587	472	516
La	53.9	49.2	61.6	36.6	51.9
Ce	109.7	98.9	122.3	75.3	107.2
Pr	12.7	11.5	14.3	8.8	12.5
Nd	52.8	48.7	57.7	37.3	54.3
Sm	10.4	10.7	11.1	8.3	10.8
Eu	3.4	3.5	3.7	2.5	3.4
Gd	8.6	9.0	9.3	6.6	8.7
Tb	1.2	1.3	1.3	1.0	1.2
Dy	6.6	6.9	6.5	5.1	6.5
Ho	1.2	1.3	1.3	1.1	1.2
Er	2.8	3.1	2.9	2.6	2.6
Tm	0.4	0.6	0.4	0.4	0.4
Yb	2.3	3.2	2.7	1.9	2.2
Lu	0.3	0.6	0.4	0.3	0.3
Hf	6.4	6.9	6.3	4.9	6.4
Ta	5.5	5.2	5.9	3.8	5.2
Pb	3.8	4.4	3.6	3.2	3.5
Th	6.3	5.9	5.6	3.6	6.1
U	1.6	1.9	1.5	1.0	1.5
ΔNb	0.3	0.2	0.3	0.3	0.3

Basalt	BA02	BA11	BA61	BA87	BA88
Wt. %					
SiO ₂	44.27	44.19	45.4	48.68	45.6
TiO ₂	3.65	3.68	2.52	2.47	2.82
Al ₂ O ₃	13.21	15.77	13.25	14.72	14.04
Fe ₂ O ₃	13.06	13.95	12.3	12.21	12.44
MnO	0.16	0.15	0.17	0.16	0.2
MgO	10.13	5.53	10.72	7.17	9.73
CaO	9.33	11.14	9.64	9.58	8.37
Na ₂ O	2.65	2.68	3.02	3.34	3.27
K ₂ O	0.96	0.63	1.16	0.84	1.51
P ₂ O ₅	0.53	0.66	1.03	0.53	0.71
LOI	1.68	1.2	0.38	0.24	1.2
Total	99.63	99.58	99.59	99.94	99.89
Mg #	59	42	62	52	59
ppm					
Nb	33	33.1	77.9	71.1	55.8
Zr	159	162	222	147	251
Y	21.6	23.2	30	22.9	25.3
Sr	851	874	1540	504	744
Rb	24.8	12.0	41.0	48.9	32.3
Ga	19.8	4.8	19.9	21.1	19.6
Ni	196	57.5	84.4	145	187
Co	53.1	46.1	33.2	40.4	42.5
Cr	556	98.5	194	260	321
Sc	23.3	19.2	3.9	18	16.2
V	249	299	233	203	195
Cs	0.4	0.6	0.8	0.2	0.4
Ba	394	248	1095	310	437
La	25.4	27.5	63.2	26.2	42.3
Ce	56.7	51.1	127.8	147.8	81.9
Pr	7.1	7.4	15.3	6.2	9.7
Nd	32.5	32.9	63.3	26.6	40.1
Sm	7.1	7.7	12.1	6.2	7.9
Eu	2.6	2.6	4.5	2.3	2.7
Gd	6.1	6.1	9.4	5.7	6.9
Tb	0.9	0.9	1.2	0.9	1.0
Dy	4.5	4.8	6.6	4.6	5.4
Ho	0.8	1.0	1.1	0.9	1.0
Er	1.9	2.0	2.6	2.0	2.2
Tm	0.3	0.3	0.3	0.3	0.3
Yb	1.6	1.7	1.9	1.5	1.9
Lu	0.2	0.2	0.3	0.3	0.3
Hf	3.8	3.8	5.2	0.9	5.7
Ta	2.4	2.5	5.6	2.1	4.0
Pb	2.2	1.8	4.2	1.9	2.5
Th	2.0	2.3	5.7	2.3	4.6
U	0.6	0.6	1.5	0.4	1.2
ΔNb	0.3	0.3	0.5	0.3	0.2

Hawaiite	BA41	BA42	BA43	BA78	BA86
Wt.%					
SiO ₂	45.86	45.14	45.65	46.19	46.14
TiO ₂	3.57	3.19	2.86	2.88	2.54
Al ₂ O ₃	16.84	15.78	17.31	15.29	14.82
Fe ₂ O ₃	13.37	12.82	11.67	12.27	12.37
MnO	0.16	0.21	0.2	0.19	0.18
MgO	5.04	5.12	3.47	6.67	7.95
CaO	7.76	8.53	8.88	8.8	8.9
Na ₂ O	3.67	3.59	4.59	3.69	3.88
K ₂ O	1.68	1.57	1.82	1.75	1.84
P ₂ O ₅	0.66	1.83	1.25	0.83	0.75
LOI	1.08	2	1.04	1.2	0.31
Total	99.69	99.78	98.74	99.76	99.68
Mg #	41	42	39	50	54
ppm					
Nb	58	64.3	77.8	70.1	63.7
Zr	284	292	277	292	261
Y	28.5	33.6	32.5	27.6	24.5
Sr	826	1860	1154	1001	813
Rb	35.5	28.0	42.8	41.4	41.9
Ga	23.6	21.5	22.8	22.7	22.2
Ni	16.8	2.4	2.16	85	152
Co	35.3	20.5	19.6	36.3	39.2
Cr	8.2	3.4	1.59	172	245
Sc	14.5	14.3	13.9	15.6	16
V	222	155	134	197	189
Cs	0.3	1.1	0.5	0.5	0.4
Ba	540	977	587	710	695
La	46.5	60.2	61.6	55.7	50.1
Ce	165.6	59.4	122.3	108.7	96.7
Pr	11.6	16.7	14.3	12.1	11.2
Nd	47.1	72.6	57.7	50.0	45.1
Sm	9.2	14.3	11.1	9.3	8.9
Eu	2.9	5.5	3.7	3.2	2.7
Gd	7.5	11.0	9.3	7.9	7.0
Tb	1.1	1.4	1.3	1.0	1.0
Dy	6.1	7.4	6.5	6.0	5.4
Ho	1.1	1.6	1.3	1.1	1.0
Er	2.7	2.9	2.9	2.4	2.3
Tm	0.4	0.4	0.4	0.3	0.3
Yb	2.3	2.3	2.7	2.1	1.9
Lu	0.3	0.4	0.4	0.3	0.3
Hf	1.1	1.4	6.3	6.3	6.0
Ta	4.3	4.9	5.9	5.1	4.5
Pb	3.3	2.6	3.6	4.5	4.2
Th	4.6	5.0	5.6	7.1	5.7
U	1.1	1.2	1.5	1.6	1.4
ΔNb	0.1	0.2	0.3	0.2	0.2

Mugearite	BA39	BA44	BA74	BA79	BA80
Wt. %					
SiO ₂	49.47	49.04	49.97	47.59	48.07
TiO ₂	1.89	2.13	1.6	2.06	2.04
Al ₂ O ₃	17.31	16.57	15.88	15.53	15.48
Fe ₂ O ₃	11.94	10.67	13.56	12.19	13.11
MnO	0.21	0.17	0.21	0.21	0.21
MgO	2.37	3.08	2.42	2.71	3.65
CaO	6.54	7.08	5.12	6.17	6.05
Na ₂ O	4.93	4.34	4.82	4.85	4.72
K ₂ O	2.2	2.33	2.71	2.23	2.24
P ₂ O ₅	0.75	0.91	1.14	1.51	1.5
LOI	1.89	3.19	2	4	2
Total	99.5	99.51	99.43	99.05	99.07
Mg #	27	35	25	29	34
ppm					
Nb	67.3	71.1	79.9	74	75.4
Zr	409	347	432	374	406
Y	30.5	34.4	40.9	44.9	44.1
Sr	892	1061	939	762	750
Rb	53.8	48.9	41.8	32.4	30.7
Ga	26.4	24	24.5	9.7	9.8
Ni	4.4	15.4	1.5	3	2.2
Co	18.3	18.1	9.7	15.8	15.1
Cr	6.3	17.1	1.8	6.3	5.1
Sc	9.5	15.3	12.6	10.6	10.8
V	79	97.2	5.3	45.3	44.9
Cs	0.5	0.6	0.4	0.3	0.7
Ba	1064	1233	1296	658	645
La	59.4	74.2	77.3	74.1	73.3
Ce	121.1	151.7	122.3	96.5	132.2
Pr	12.9	17.9	20.4	19.2	19.4
Nd	50.7	73.6	84.8	85.2	82.9
Sm	9.4	13.1	15.3	16.3	16.4
Eu	3.2	4.6	5.8	5.2	5.2
Gd	7.7	11.0	12.0	13.7	13.6
Tb	1.2	1.5	1.7	1.8	1.9
Dy	6.0	7.3	8.4	9.7	9.8
Ho	1.2	1.4	1.6	1.9	1.9
Er	3.0	3.1	3.7	4.0	3.9
Tm	0.4	0.5	0.5	0.6	0.6
Yb	2.6	2.9	3.2	3.5	3.5
Lu	0.4	0.5	0.5	0.5	0.5
Hf	1.2	8.6	1.6	8.9	9.2
Ta	5.0	5.5	5.8	8.9	9.2
Pb	5.5	7.	5.0	4.8	5.0
Th	7.4	6.1	7.1	6.5	6.4
U	1.8	1.7	1.6	1.6	1.5
ΔNb	-0.08	0.1	0.1	0.2	0.1

$Mg\# = 100 * (MgO/40.31) / (MgO/40.31 + Fe_2O_3 * 0.8998 / (71.85 * (1 - 0.15)))$, assuming $Fe_2O_3 / (Fe_2O_3 + FeO) = 0.15$. Recalculated to 100% anhydrous; total iron is expressed as Fe_2O_3 ; LOI: Loss on ignition.

Table 3: K-Ar ages obtained on some mafic rocks from the Bamenda area.

Sample	Rock	K ₂ O (%)	⁴⁰ Ar* (10 ¹⁵ at/g)	⁴⁰ Ar* (10 ⁻⁶ cm ³ /g)	Age (Ma)	Error (1s)	Laboratory
BA61	Basalt	1.19	0	0	0	0	CRPG (Nancy)
BA31	Basanite	1.65	0.0006	0.021	0.4	0.1	CRPG (Nancy)
BA10	Basanite	1.52	0.00189	n.a	1.2	0.2	CRPG (Nancy)
BA24	Basanite	1.91	0.0021	0.079	1.3	0.1	CRPG (Nancy)
BA73	Basanite	2.016	0.0076	0.282	5.5	0.4	CRPG (Nancy)
BA37	Basanite	1.31	0.0086	0.322	7.6	0.2	CRPG (Nancy)
BA68	Basanite	1.72	0.0124	0.462	8.3	0.3	CRPG (Nancy)
BA42	Hawaiite	1.6	0.0236	0.877	16.9	0.3	CRPG (Nancy)
BA60	Hawaiite	1.91	0.03525	n.a.	17.6	0.3	CRPG (Nancy)
BA60	Duplicate		0.0351	n.a.	17.4	0.8	CRPG (Nancy)
Sample	Rock	K ₂ O (%)	⁴⁰ Ar* (%)	⁴⁰ Ar* (10 ⁻¹² mol/g)	Age (Ma)	Error (1s)	Laboratory
BA87	Basalt	0.63	8.48	4.011	4.4	1.4	Univ. Queensland(Australia)

The following decay constants were used: $^{40}\lambda_e = 0.581 \times 10^{-10} \text{ a}^{-1}$ and $^{40}\lambda_\beta = 4.962 \times 10^{-10} \text{ a}^{-1}$. n.a.: not available.

Table 4. Sr-Nd-Pb isotopic compositions of the rocks from Bamenda Mountains. na: not analysed.

Sample	Rock type	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵ Nd	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
BA10	Basanite	0.703078	0.512895	5.01	na	na	na
BA24	Basanite	0.703138	0.512852	4.17	na	na	na
BA46	Basanite	0.703352	0.512890	4.92	20.217	15.645	39.729
BA47	Basanite	0.703338	0.512886	4.84	20.250	15.645	39.751
BA52	Basanite	0.703505	0.512812	3.39	na	na	na
BA68	Basanite	0.703702	0.512797	3.10	19.417	15.616	39.302
BA84	Basanite	0.703767	0.512828	3.71	19.591	15.536	38.537
BA11	Basalt	0.703340	0.512860	4.33	na	na	na
BA27	Basalt	0.703935	0.512757	2.32	na	na	na
BA65	Basalt	0.704120	na	na	na	na	na
BA87	Basalt	0.703375	0.512898	5.07	na	na	na
BA88	Basalt	0.703070	0.512911	5.33	na	na	na
BA06	Hawaiite	0.703029	0.512927	5.64	na	na	na
BA41	Hawaiite	0.703557	0.512733	1.85	18.198	15.461	37.805
BA42	Hawaiite	0.704133	0.512743	2.05	na	na	na
BA56	Hawaiite	0.703951	0.512795	3.06	19.131	15.568	39.058
BA60	Hawaiite	0.704243	0.512616	-0.43	17.317	15.422	37.114
BA66	Hawaiite	0.703949	0.512761	2.40	Na	na	na
BA77	Hawaiite	0.703945	0.512820	3.55	19.494	15.601	39.682
BA78	Hawaiite	0.703945	na	na	18.921	15.536	38.537
BA86	Hawaiite	0.703387	0.512809	3.34	na	na	na
BA39	Mugearite	0.704031	0.512755	2.28	na	na	na
BA79	Mugearite	0.703803	0.512769	2.56	na	na	na

Representative microprobe analyses of olivine from selected lavas from Bamenda area.

Sample	BA10			BA37		
	Basanite					
	r	c	xenocryst	m	c	r
SiO ₂	36.65	39.55	40.84	37.35	39.81	38.63
TiO ₂	0.07	0.00	0.09	0.05	0.00	0.05
Al ₂ O ₃	0.00	0.03	0.03	0.00	0.08	0.00
FeO	30.23	16.84	9.21	28.43	15.17	23.56
MnO	0.72	0.23	0.10	0.54	0.08	0.39
MgO	32.11	42.52	48.83	33.62	44.06	37.03
CaO	0.50	0.11	0.10	0.35	0.25	0.47
NiO	0.00	0.28	0.32	0.00	0.07	0.00
Cr ₂ O ₃	0.02	0.03	0.06	0.00	0.00	0.00
Total	100.30	99.59	99.58	100.34	99.52	100.13
Formulae on basis of 4 oxygens						
Si	0.99	1.01	1.00	1.00	1.00	1.01
Fe ²⁺	0.68	0.36	0.19	0.64	0.32	0.51
Mn	0.02	0.00	0.00	0.01	0.00	0.01
Mg	1.30	1.61	1.79	1.34	1.66	1.44
Ca	0.01	0.00	0.00	0.01	0.01	0.01
Ni	0.00	0.01	0.01	0.00	0.00	0.00
X _{Mg}	0.65	0.82	0.90	0.68	0.84	0.74
Fo	65.44	81.82	90.43	67.83	83.81	73.70

r = rim; c= core; m= microlite