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1 Plio-Pleistocene intra-plate magmatism from the southern Sulu  
2 Arc, Semporna peninsula, Sabah, Borneo: Implications for high-  
3 Nb basalt in subduction zones

4  
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## 32 **Abstract**

33 New analyses of major and trace element concentrations and Sr, Nd and Pb isotopic ratios  
34 are presented for Plio-Pleistocene basalts and basaltic andesites from the Semporna  
35 peninsula in Sabah, Borneo, at the southern end of the Sulu Arc. Depletion of high field  
36 strength elements (HFSE), which is characteristic of many subduction-related magmatic  
37 suites, is present in more evolved Semporna rocks but is associated with radiogenic Sr and  
38 Pb, and less radiogenic Nd isotopic ratios and results from contamination of mafic melt by,  
39 possibly ancient, crustal basement. The most mafic lavas from Semporna, and elsewhere in  
40 the Sulu Arc, display no HFSE depletion relative to other elements with similar compatibility.

41 High-Nb basalt from Semporna formed when mantle resembling the source of Ocean Island  
42 Basalt (OIB) upwelled into lithospheric thin spots created during earlier subduction. This  
43 mantle did not experience enrichment by fluids or melt derived from subducted crust. The  
44 presence of similar lavas throughout the Sulu Arc and around the South China Sea suggests  
45 that the OIB-like component resides in the convecting upper mantle. Depletion of light rare  
46 earth elements, with respect to other incompatible elements, throughout the Sulu Arc could  
47 result from melt-mantle interaction during magma transport through the lithosphere. Such  
48 depletion is absent in suites from the South China Sea, where magma probably migrated  
49 along large, lithosphere penetrating structures.

50 Semporna high-Nb basalts are not associated with adakitic magmatism which is a frequent,  
51 but not ubiquitous, association in some active subduction zones. Both geochemical  
52 signatures are developed early in the history of a melt pulse, either in the source (high-Nb  
53 basalt) or during deep differentiation (adakite). Preservation of these distinctive geochemical  
54 signatures is favoured in settings that minimise (i) interaction with other, more copious melt  
55 types, or (ii) subsequent differentiation in the shallow crust. Where found, the high-Nb basalt  
56 – adakite association is a result of transport through favourable lithospheric conditions and  
57 not due to any link between their mantle sources.

58 Keywords: High-Nb basalt; Nb-enriched basalt; Sabah; Borneo; subduction; OIB;  
59 magmatism

## 60 **1. Introduction**

61 Subduction is an important process in generating new crust at the present time and may  
62 have played a crucial role in generating continental crust throughout much of Earth history

63 (Rudnick, 1995). Understanding subduction, and the crust that it produces, requires an  
64 understanding of spatial and temporal variations of magmatic products generated both within  
65 individual subduction zones and between different subduction zones. A striking feature of  
66 magmatism in modern volcanic arcs is the marked depletion of high field strength elements  
67 (HFSE, such as Nb, Zr and Ti) relative to other elements with similar compatibilities. This  
68 depletion is thought to result from differential transport of HFSE compared to other elements  
69 during recycling from the slab to the mantle wedge (Thirlwall et al., 1994). Many subduction-  
70 related basalts also possess low absolute concentrations of Nb.

71 Although common, relative depletion of HFSE is not ubiquitous in arc magmatism. Several  
72 subduction zones have generated basaltic magma in which Nb, and most other incompatible  
73 elements, are abundant and in which there is negligible depletion of HFSE relative to  
74 elements with similar compatibility. Reagan and Gill (1989) introduced the term “high-Nb  
75 basalt” to describe such rocks from the Costa Rican volcano Turrialba that contain 36ppm  
76 Nb which is not depleted relative to Light Rare Earth Elements (LREE), such as La, or Large  
77 Ion Lithophile Elements (LILE). For example, the value of  $(\text{Nb/La})_n$ , the Nb/La ratio  
78 normalised to the value for normal mid-ocean ridge basalt (N-MORB), is 0.91 in Turrialba  
79 high-Nb basalt, while values range from 0.33 to 0.42 in more typical arc-related magmatism  
80 from the same volcano. Sajona et al. (1994) subsequently used the term “Nb-enriched  
81 basalt” for basaltic rocks from Mindanao, the Philippines, containing 4-16ppm Nb and with  
82  $(\text{Nb/La})_n$  ranging between 0.72 and 1.41.

83 Two main mechanisms have been proposed that might generate high-Nb basalt in these and  
84 other convergent margins. Reagan and Gill (1989) concluded that incompatible trace  
85 element enrichment is inherited from small-degree partial melts of an Ocean Island Basalt  
86 (OIB)-like source, which then interact with high-degree partial melts of depleted upper  
87 mantle. The OIB melts are undersaturated in rutile because they carry reduced C-O-H fluids  
88 and so Nb is not depleted relative to elements with similar compatibility. This contrasts with  
89 contemporaneous, presumed rutile-saturated, calc-alkaline magmatism at the same volcanic  
90 centres. Variations on this theme, with or without contributions from subducted crust and  
91 sediment, have been proposed for several locations (Storey et al., 1989; Leeman et al., 1990  
92 and 2005; Richards et al., 1990; Petrone et al., 2003; Castillo et al., 2002 and 2007; Castillo,  
93 2008; Petrone and Ferrari, 2008).

94 An alternative group of models arises from the observation that several high-Nb basalt suites  
95 occur in subduction zones where the subducting plate is young and, therefore, hot (Defant et

96 al., 1992). Based on the premise that young subducted slabs are prone to melting (Defant  
97 and Drummond, 1990) and on the presence of putative slab melt magmatism in association  
98 with some high-Nb basalt occurrences, Defant et al. (1992) proposed that high-Nb basalt  
99 may be produced from mantle into which metasomatic, Nb-rich amphibole has been  
100 introduced by slab melt. This model has subsequently been applied to several high-Nb  
101 basalt occurrences (Sajona et al., 1994 and 1996; Kepezhinskas et al., 1995, 1996 and  
102 1997; Escuder Viruete et al., 2007; Gómez-Tuena et al., 2007).

103 In this contribution we discuss the origin of Plio-Pleistocene high-Nb basalt magmatism from  
104 the Semporna peninsula of Sabah, Malaysia in northeastern Borneo. This site lies at the  
105 southern end of the Sulu Arc, an arcuate band of magmatism extending south-eastwards  
106 from the Zamboanga peninsula in western Mindanao through the Sulu Islands, such as  
107 Basilan and Jolo, towards NE Borneo (Fig. 1a). There is field and petrological evidence  
108 which suggests that the Sulu Arc produced subduction-related magmatism during the  
109 Miocene (Section 2.1). A deep trench (> 4800m) with high heat-flow lies to the north of the  
110 arc but Hamilton (1979) considered that this trench does not represent on-going subduction.  
111 There is little significant seismic activity currently associated with the Sulu Arc and  
112 tomographic imaging provides no evidence for a subducted slab beneath the islands  
113 (Spakman and Bijward, 1998; Rangin et al., 1999). Thus, although the term arc is  
114 appropriate to the bathymetry of the system it should not be used to infer that subduction is  
115 active, or that Plio-Pleistocene magmatism was caused by subduction-related processes.

116 We compare high-Nb magmatism from Sabah to magmatism in the rest of the Sulu Arc and  
117 to magmatic suites found elsewhere in SE Asia to investigate the nature of the mantle  
118 source and the lithosphere beneath Sabah. Then we discuss the implications of our findings  
119 for understanding mechanisms that might generate high-Nb basalt.

## 120 **2. Setting and Samples**

### 121 *2.1 Mio-Pliocene Magmatism*

122 Eurasia's eastern margin has interacted with the Pacific Plate throughout the Cenozoic  
123 generating a complex assemblage of plate fragments (Fig. 1a). The basement of Sabah was  
124 produced through accretion of Cretaceous ophiolitic fragments to the continental core of the  
125 island (Hall, 2002). From the Paleogene until the Early Miocene, southward-directed  
126 subduction of the proto-China Sea produced an accretionary margin in northern Borneo. The  
127 latter stages of this convergence occurred as the South China Sea Basin was opening to the

128 north. K-Ar analyses obtained Middle to Late Miocene (12.9-9 Ma, Rangin et al., 1990;  
129 Bellon and Rangin, 1991) or Middle Miocene (18.8-14.4 Ma, Swauger et al., 1995) ages for  
130 Neogene magmatism in the Semporna and neighbouring Dent peninsulas, although these  
131 dates are uncertain because they were determined on whole rock samples that may have  
132 been subject to tropical weathering. The petrography and geochemistry of this magmatism is  
133 consistent with genesis in an island arc (Bellon & Rangin, 1991; Hutchison et al., 2000,  
134 Chiang, 2002) but the lack of tomographic evidence for dipping slabs, either modern or  
135 ancient (Spakman and Bijward, 1998; Rangin et al., 1999), has complicated efforts to  
136 determine the polarity of Neogene subduction. Hall (2002) used the geology of Sabah to  
137 infer that this subduction was directed towards the northwest. Chiang (2002) investigated  
138 this further by examining incompatible trace element ratios of Neogene arc magmatism  
139 throughout SE Sabah and also concluded that Celebes Sea crust was subducted beneath  
140 the Sulu Arc towards the northwest.

## 141 *2.2 Plio-Pleistocene Magmatism*

142 The youngest phase of magmatism in Sabah is the subject of this work. Plio-Pleistocene  
143 basalt and basaltic andesite lavas, cinder cones and occasional dykes are found at Tawau  
144 and Mostyn on the Semporna Peninsula (Fig. 1). At Tawau eruptions occurred through  
145 cinder cones while the Mostyn eruptions mainly occurred along N130°E-trending fissures.  
146 The ages of these rocks are poorly known. Lim and Hen (1985) suggested ages of 27 Ka or  
147 younger, while Rangin et al. (1990) obtained whole rock K-Ar dates of 2.8-3.1 Ma. However,  
148 Bellon and Rangin (1991) concede that the K-Ar data remain suspect, concluding that  
149 volcanism injected along the fissures is probably very young and that these faults are still  
150 active.

151 Plio-Pleistocene lavas from Tawau are aphyric to moderately (<20%) porphyritic basalts and  
152 basaltic andesites. Fresh plagioclase and olivine are the most abundant phenocryst phases  
153 in the basalts with magnetite and clinopyroxene occurring as minor phases in the  
154 groundmass. In the basaltic andesites, clinopyroxene is the most abundant phenocryst  
155 phase. Plagioclase is also the most abundant phenocryst phase in the Mostyn suite. Large,  
156 fresh phenocrysts of olivine are found in the most basic rocks and small orthopyroxene  
157 phenocrysts are the most common mafic phenocryst in more silicic rocks, although the total  
158 phenocryst content is particularly low in the latter. Clinopyroxene is a minor phase in the  
159 matrix of most Mostyn lavas.

### 160 3. Techniques

161 XRF analyses were performed using a Philips PW1480 XRF spectrometer at Royal  
162 Holloway, University of London. LOI was determined by heating the pre-dried sample at  
163 1100°C for 20 minutes. Major element concentrations were analysed on fused discs of pre-  
164 dried sample mixed with pre-dried La<sub>2</sub>O<sub>3</sub> Johnson-Matthey Spectroflux 105 (ratio sample:flux  
165 = 1:6). Trace element (Ni, Cr, V, Sc, Cu, Zn, Cl, Ga, Pb, Sr, Ba, Zr, Nb, Th, Y, La, Ce and  
166 Nd) concentrations were determined on pressed power pellets with matrix corrections based  
167 on major element compositions. Reproducibility (2sd of six replicate preparations) of XRF  
168 data is reported in Table 1; based on 25-35 international standards accuracy is comparable  
169 for major elements and for trace elements where these have been analysed by isotope  
170 dilution.

171 Sr and Nd isotopic analyses were conducted at the Arthur Holmes Isotope Geochemistry  
172 Laboratory at the University of Durham using a ThermoElectron Neptune multi-collector ICP-  
173 MS system. Details of the operating procedures and instrument configuration are given in  
174 Handley et al. (2007). Measured values for the NBS 987 and J&M standards  $\pm 2SD$  error  
175 during the same runs as the Semporna samples were  $0.710270 \pm 18$  (n=11) and  $0.511101 \pm 5$   
176 (n=15), respectively. Data are reported relative to NBS 987 and J&M standard values of  
177 0.71024 (Thirlwall, 1991) and 0.511110 (Royse et al., 1998), respectively. Total procedural  
178 blanks for Sr and Nd were determined by ICP-MS on a PerkinElmer ELAN 6000 quadrupole  
179 ICP-MS system at Durham University and were below 1.2 ng for Sr and 219 pg for Nd.  
180 These values are considered insignificant in relation to the quantity of Sr and Nd typically  
181 processed from Semporna rocks.

182 Pb isotope ratios were determined at the Scripps Institution of Oceanography following the  
183 procedure described in Janney and Castillo (1996; 1997). Rock powders were dissolved with  
184 a double-distilled, 2:1 mixture of concentrated HF:HNO<sub>3</sub> acid in Teflon beakers. Lead was  
185 separated from sample solutions using small ion exchange columns in an HBr medium and  
186 its isotopes were measured using a 9-collector, Micromass Sector 54 thermal ionization  
187 mass spectrometer. Lead isotopes were fractionation corrected using the isotope values of  
188 NBS 981 relative to those of Thirlwall (2000). Analytical uncertainties based on repeated  
189 measurements of standards are  $\pm 0.008$  for  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $\pm 0.030$  for  
190  $^{208}\text{Pb}/^{204}\text{Pb}$ . Routine analytical blank was generally <30 pg of Pb.

#### 191 4. Results

192 Plio-Pleistocene lavas from Tawau and Mostyn display limited ranges of SiO<sub>2</sub> (49.44 to  
193 56.56 wt.%) and MgO (4.36 to 7.66 wt.%). The Mostyn group are distinct from Tawau lavas  
194 in having lower K<sub>2</sub>O and P<sub>2</sub>O<sub>5</sub>, and higher Fe<sub>2</sub>O<sub>3</sub> and TiO<sub>2</sub> at any value of MgO (Fig. 2). For  
195 most major elements there is a significant amount of scatter at any MgO content. No  
196 correlations were found between major element concentrations and the modal abundance of  
197 any phenocryst phase. For the suite, as a whole, there is a general increase in SiO<sub>2</sub> with  
198 decreasing MgO, but most major elements show no simple variation with differentiation. This  
199 is also apparent for incompatible trace elements (Fig. 3). The Tawau lavas can be divided  
200 into four sub-groups (PP1 to PP4) for which, at any concentration of MgO, the  
201 concentrations of K<sub>2</sub>O, P<sub>2</sub>O<sub>5</sub>, Sr, Nb, Rb and Ba all decrease in the order PP4 > PP3 > PP2  
202 > Mostyn (Figs. 2 and 3). For Pb, Th, La, Ce and Nd a similar order exists but with PP2 lying  
203 between PP3 and PP4. The PP1 group has a very restricted range in MgO but frequently  
204 has incompatible element concentrations similar to, or lying on an extension of, the PP2  
205 trend.

206 Concentrations of Ni (Fig. 3a) and Cr decrease with MgO as would be expected in magmas  
207 produced by differentiation of basaltic parents. In contrast, the majority of incompatible  
208 elements in the Tawau sub-groups and Mostyn lavas show behaviour that is not consistent  
209 with their expected compatibility in basaltic magma. The concentrations of these elements  
210 should increase as MgO falls but concentrations of K<sub>2</sub>O, P<sub>2</sub>O<sub>5</sub> and most other incompatible  
211 elements show relatively little variation or pronounced decreases with decreasing MgO  
212 (Figs. 2 and 3). This effect is particularly notable for P<sub>2</sub>O<sub>5</sub> in the PP3 and Mostyn groups,  
213 which could indicate crystallisation of apatite, however, Semporna lavas are more mafic than  
214 would be suitable for apatite saturation at these P<sub>2</sub>O<sub>5</sub> concentrations in basalt or alkali basalt  
215 (DeLong and Chatelain, 1990; Busà et al., 2002).

216 With respect to N-MORB, Semporna Plio-Pleistocene lavas contain high concentrations of  
217 the majority of incompatible elements (Fig. 4). As already noted, Tawau samples possess  
218 higher concentrations of most trace elements for any particular MgO content. The patterns  
219 for the most mafic rock from Tawau (SBK 13) and Mostyn have the smoothest patterns, in  
220 which HFSE display negligible depletion relative to neighbouring elements. This contrasts  
221 with typical subduction-related magmas, including Mio-Pliocene lavas from Tawau (Chiang,  
222 2002). Negative relative Nb anomalies become increasingly apparent in more evolved rocks  
223 (Fig. 5). Concentrations of all LILE are elevated with respect to MORB. This is most



224 apparent for Pb, which displays prominent positive anomalies compared to neighbouring  
225 elements in the MORB-normalised plot, but Pb is no more enriched than other LILE with  
226 respect to MORB (Fig. 4). Overall, the Semporna lavas possess higher concentrations of  
227 LILE and LREE than MORB and more closely resemble OIB than typical subduction-related  
228 magmatism, although LILE/LREE ratios are lower than in OIB. The most mafic Semporna  
229 lavas have HFSE/LILE and HFSE/LREE ratios comparable to or even higher than OIB.

230 Semporna Plio-Pleistocene lavas possess wide ranges in  $^{87}\text{Sr}/^{86}\text{Sr}$  (0.704092 to 0.706291),  
231  $^{143}\text{Nd}/^{144}\text{Nd}$  (0.512846 to 0.512491) and Pb isotope ratios ( $^{206}\text{Pb}/^{204}\text{Pb}$ ; 18.528 to 18.871,  
232  $^{207}\text{Pb}/^{204}\text{Pb}$ ; 15.566 to 15.667 and  $^{208}\text{Pb}/^{204}\text{Pb}$ ; 38.598 to 39.116). Taking the Semporna  
233 lavas as a single suite the more evolved lavas tend to possess higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and Pb  
234 isotopic ratios and lower  $^{143}\text{Nd}/^{144}\text{Nd}$ . Despite the limited number of analyses this statement  
235 is also true for each site, with the Mostyn lavas offset to slightly higher  $^{87}\text{Sr}/^{86}\text{Sr}$  and Pb, and  
236 lower  $^{143}\text{Nd}/^{144}\text{Nd}$  at similar MgO (Fig. 6). Despite these offsets, co-variations between  
237 isotope ratios of different elements are particularly well defined for the suite as a whole. The  
238 most mafic lava has isotopic ratios that lie within the field of Indian Ocean MORB and that  
239 resemble values for several other small-volume volcanic provinces in SE Asia, but the more  
240 evolved lavas extend well beyond the field of Indian MORB (Fig. 7).

## 241 5. Discussion

242 The most mafic Semporna lava contains 7.66 wt.% MgO, which is close to the upper range  
243 of MgO contents in high-Nb basalts from other locations (8-9 wt.%), and is likely to provide a  
244 good estimate of the composition of parental magma. Although, SBK13 has a smooth trace  
245 element pattern (Fig. 4), mild to moderate Nb depletion ( $(\text{Nb}/\text{K})_n < 1$ ) is apparent in many  
246 Semporna rocks, which might indicate a role for subduction-modified mantle. To understand  
247 the extent to which the relative Nb depletion of Semporna lavas is inherited from the mantle  
248 source it is necessary to determine how differentiation has affected trace elements.

### 249 5.1 Nb depletion of Semporna Plio-Pleistocene magma during differentiation

250 It is unlikely that the geochemical variations within the Semporna Plio-Pleistocene lavas  
251 result from fractional crystallisation of a uniform parental magma composition. First,  
252 concentrations of several elements that are usually incompatible during crystallisation of  
253 basaltic magma decrease with MgO. In all of the sub-groups  $\text{P}_2\text{O}_5$ , Nb and Sr decrease  
254 strongly from basalt to basaltic andesite. In the Mostyn group the other LILE and La also  
255 show the same effect (Fig. 3). Second, variations in Nb, K and La suggest unusual

256 behaviour between HFSE, LILE and LREE. These elements have similar compatibilities in  
257 basaltic magma, therefore (Nb/La)<sub>n</sub> and (Nb/K)<sub>n</sub> should change little as basalt differentiates  
258 to basaltic andesite in a closed system. However, with the exception of relatively high Nb/La  
259 in two PP1 samples, both ratios become lower as MgO decreases (Fig. 5). Third, like Nb/La  
260 and Nb/K, the isotopic ratios of Sr, Nd and Pb should not vary in a suite of lavas produced by  
261 fractional crystallisation of uniform parental magma. In the Semporna lavas there are large  
262 ranges in each of these ratios, which change from the most to least evolved rocks (Fig. 6  
263 and 7).

264 The strong inter-isotope correlations suggest that two main components are involved at  
265 Semporna (Fig. 7). Magma mixing, either between basic and evolved melts, or two mafic  
266 melts with different sources, is considered unlikely, since no petrographic evidence was  
267 found to indicate such a process. Therefore, we conclude that mafic, mantle-derived magma  
268 was contaminated by crust possessing high <sup>87</sup>Sr/<sup>86</sup>Sr and Pb isotope ratios and low  
269 <sup>143</sup>Nd/<sup>144</sup>Nd during differentiation from basalt to basaltic andesite. It is difficult to determine  
270 the nature of the contaminant because there are very few analyses of the compositions, and  
271 particularly isotopic ratios, of basement lithologies in Sabah. However, a number of  
272 reasonable inferences can be made. Mio-Pliocene arc magmatism from Sabah is  
273 characterised by Sr and Pb isotope ratios that are too low and by Nd isotope ratios that are  
274 too high to be the contaminant (Figs. 6). Similarly, the Mesozoic ophiolitic basement of  
275 Sabah, which is most likely to be comprised of fragments of oceanic crust resembling Indian  
276 Ocean MORB, would possess low <sup>86</sup>Sr/<sup>86</sup>Sr and high <sup>143</sup>Nd/<sup>144</sup>Nd (Omang and Barber, 1996;  
277 Weis and Frey, 1996).

278 In Sabah the only exposures of rocks derived from the lower crust are granite bodies. Mount  
279 Kinabalu, in north Sabah, is a composite granite that was intruded over a short period during  
280 the Miocene (Cottam et al., in press). Unfortunately, sufficient isotopic data do not exist to  
281 conduct detailed modelling of the effects of contamination by this material. However, the  
282 <sup>86</sup>Sr/<sup>86</sup>Sr range, from 0.706364 to 0.707832 (Chiang, 2002), suggests that material of this  
283 type would not be a suitable contaminant. The lower end of the range would require more  
284 than 95% contamination of SBK13 to produce the highest <sup>86</sup>Sr/<sup>86</sup>Sr in the Semporna suite  
285 while the upper end of the range would require 60% contamination. In both cases, such  
286 levels of contamination would produce magma that was more silicic than the most evolved  
287 basaltic andesite. Assimilation of such material can be examined further by using an  
288 analogous granitic body from the island of Palawan (Fig. 1). For reasonable values of r (the

289 ratio of mass assimilated to mass crystallised), assimilation of Capoas granite during  
290 fractional crystallisation of SBK13 fails to produce an array resembling the Semporna dataset  
291 (model LC 1 in Fig. 8). Bulk mixing of SBK13 with magma resembling the Capoas granite  
292 might achieve a closer fit to the isotopic and major element characteristics of the Semporna  
293 suite but would still require up to 40% contamination by the granitic component. As  
294 discussed above, there is no evidence of magma mixing in the Semporna rocks, which  
295 should be readily observed for such extensive contamination.

296 East Indonesian sediments (Vroon et al., 1995) may contain components that resemble the  
297 crustal fragments which were incorporated into SE Asian lithosphere. Therefore, we have  
298 also explored assimilation with fractional crystallisation (AFC) models using these sediments  
299 as crustal contaminants. Despite suitable Sr and Pb characteristics even the most extreme  
300 East Indonesia sediment composition does not possess sufficiently low  $^{143}\text{Nd}/^{144}\text{Nd}$  to  
301 provide an appropriate contaminant (model EIS 1, Fig. 8). Increasing  $r$ , even to the relatively  
302 high value of 0.85, does not produce a fit to the data (model EIS 2) and reduces the amount  
303 of differentiation to the extent that there would virtually no change in major element  
304 chemistry of the magma.

305 Due to the restricted range of MgO contents in the Semporna lavas the contaminant requires  
306 very low  $^{143}\text{Nd}/^{144}\text{Nd}$ . Good fits to the Semporna dataset can be achieved for AFC models  
307 using assimilants with the isotopic characteristics of Archean crustal rocks via the moderate  
308 extents of crystallisation required to differentiate from basalt to basaltic andesite (models AC  
309 1 and AC 2, Fig. 8). Although rocks of this age are not exposed in Sabah, Palaeoproterozoic  
310 ages have been determined for detrital zircons from the Crocker Formation in northern  
311 Borneo, for which van Hattum et al. (2006) postulated a local origin, and for inherited zircon  
312 crystals in the Kinabalu granite (Cottam et al., in press). Therefore, we postulate the  
313 Semporna crust contains Archean domains. Continental fragments may have been  
314 embedded in the Mesozoic oceanic lithosphere now forming the ophiolitic basement of  
315 Sabah, or may have been incorporated during northward and westward dispersion of  
316 continental material derived from the leading edge of the Australian continent as it interacted  
317 with SE Asia during the Cenozoic (Hutchison et al., 2001; Hall, 2002). van Leeuwen et al.  
318 (2007) recently proposed such an origin for the Malino complex, northwest Sulawesi, where  
319 Archean inherited zircons have been discovered.

## 320 5.2 Semporna parental magma without Nb depletion

321 Differentiation is the primary control on the extent of Nb depletion in Semporna lavas  
322 (Section 5.1). More specifically, Nb concentrations are lower and Nb-depletion, relative to  
323 other elements, is more marked in rocks that have experienced greater amounts of crustal  
324 contamination. The Nb contents of the most mafic basalts suggest that all magma left the  
325 mantle containing sufficient Nb to be classed as high Nb-basalt but during subsequent  
326 differentiation Nb, and several other elements, were diluted as many melt batches effectively  
327 evolved into Nb-enriched basalt (Fig. 3g). Therefore, classifying these rocks as high-Nb  
328 basalt or Nb-enriched basalt has no significance for source characteristics or processes; it is  
329 simply a function of the extent to which the melts differentiated.

330 Semporna lavas define single, coherent arrays when different isotopic ratios are compared  
331 with one another (Fig. 7) suggesting that there is only minor variation in the isotopic  
332 compositions of the mantle source and of the contaminant. These arrays are also consistent  
333 with a restricted range in Sr/Nd ratios in the parental magma. Like Nb/K and Nb/La (Fig. 5),  
334 many trace element ratios show less variation towards the more mafic end of the  
335 compositional range and converge on the values in SBK13. This suggests that differentiation  
336 was responsible for generating much of the heterogeneity in both isotopic ratios and  
337 incompatible element ratios between different members of the suite.

338 The most mafic lavas from Tawau and Mostyn display sub-parallel incompatible trace  
339 element patterns suggesting that their mantle sources did not possess relative depletion of  
340 Nb (Fig. 4). These patterns are very similar to those of parental magma in the rest of the  
341 Sulu Arc (Fig. 9a). The Sr, Nd and Pb isotope ratios of the least evolved northern, central  
342 and southern Sulu suites also converge on similar values suggesting shared sources.  
343 Castillo et al. (2007) demonstrated that northern and central Sulu Arc lavas possess isotopic  
344 ratios similar to those of basalts from the Scarborough Seamounts and Reed Bank in the  
345 South China Sea (Fig. 1a). This similarity extends to Quaternary magmatism from Hainan  
346 Island on the northern margin of the South China Sea (Figs. 7 and 9b). The three South  
347 China Sea suites possess inter-element ratios very similar to OIB. The Sulu Arc suites also  
348 show OIB-like patterns, with the exception of relative depletions in LREE, Sr and P (Fig. 9a).

### 349 5.3 Source of Semporna Plio-Pleistocene magmatism

350 Two main mechanisms have been proposed to explain how the mantle sources of high-Nb  
351 magmatism might develop. One involves metasomatism of mantle by partial melts from  
352 subducted crust (Defant et al., 1992) and, as such, implies an intrinsic role for subduction in  
353 producing such sources. The other advocates an enriched mantle resembling the source of  
354 OIB (Reagan and Gill, 1989), so is independent of subduction. Resolving which mechanism  
355 is responsible for producing high-Nb sources has important implications for understanding  
356 the dynamics of the subduction zones in which they occur.

#### 357 5.3.1 Mantle metasomatism by partial melt from subducted basalt

358 Isotope ratios can be used to test whether partial melts derived from subducted lithosphere  
359 have metasomatised the Semporna mantle. Chiang (2002) examined variations in  
360 incompatible trace element ratios of Neogene magmatism across the Semporna and Dent  
361 peninsulas and concluded that the subducted slab was Celebes Sea oceanic lithosphere.  
362 Other studies have favoured the Sulu Sea as the source of the slab (e.g. Castillo et al.,  
363 2007) but the two basins are floored by basalt with similar isotopic compositions (Fig. 7) and  
364 so the distinction is irrelevant for the purpose of conducting this test. Isotopically distinctive  
365 basalt has been recovered from the Cagayan Ridge in the Sulu Sea, but the petrology and  
366 geochemistry of these rocks indicate that this bathymetric high originated as a volcanic arc  
367 (Bellon and Rangin, 1991; Spadea et al., 1991 and 1996). Features of this type have a lower  
368 probability of being subducted than the oceanic lithosphere of the adjacent basins.  
369 Therefore, the Cagayan Ridge samples are unlikely to represent a feasible slab melt  
370 composition.

371 If Semporna Plio-Pleistocene basalt originated in mantle that was metasomatised by partial  
372 melts from subducted lithosphere then (i) the most mafic Semporna lavas should possess  
373 isotope ratios closest to the compositions of Sulu or Celebes ocean floor basalt, and (ii) the  
374 Semporna isotopic arrays should trend towards that field. Some of the Semporna isotopic  
375 arrays do project back towards Sulu-Celebes compositions (e.g.  $^{143}\text{Nd}/^{144}\text{Nd}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$   
376 and  $^{208}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$ ). However, the  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  and  
377  $^{143}\text{Nd}/^{144}\text{Nd}$  versus  $^{207}\text{Pb}/^{204}\text{Pb}$  arrays clearly trend outside the range of Celebes and Sulu  
378 oceanic basalts and would infer a mantle with considerably higher  $^{206}\text{Pb}/^{204}\text{Pb}$  at the  
379 measured  $^{207}\text{Pb}/^{204}\text{Pb}$  or  $^{143}\text{Nd}/^{144}\text{Nd}$  than the putative slab compositions (Fig. 7).  
380 Furthermore, the most mafic Semporna lava, which has experienced negligible

381 contamination by crust (Section 5.1), lies significantly outside the Sulu-Celebes range for  
382  $^{87}\text{Sr}/^{86}\text{Sr}$ ,  $^{206}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$ .

383 There is good evidence that subduction occurred beneath the Sulu Arc during the Miocene  
384 but the South China Sea sites lie as much as 1500km from Sulu and have not experienced  
385 recent subduction (Fig. 1a). It is extremely unlikely that metasomatism by partial melts of  
386 different subducted slabs at different times could yield sources with similar trace element  
387 and isotopic ratios beneath the Sulu Arc, Hainan Island, Scarborough Seamounts and Reed  
388 Bank (Figs. 7 and 9b). Therefore, we conclude that the source of Semporna Plio-Pleistocene  
389 lavas did not contain a significant contribution from partially-melted subducted lithosphere.  
390 Castillo et al. (2007) reached a similar conclusion for the central and northern Sulu Arc.

### 391 5.3.2 *Intra-plate (OIB) mantle source*

392 The least contaminated Semporna lava resembles many OIB in possessing high  $^{87}\text{Sr}/^{86}\text{Sr}$   
393 and Pb isotope ratios and low  $^{143}\text{Nd}/^{144}\text{Nd}$  relative to the source of MORB (Fig. 7). Like  
394 magmatism in the South China Sea (Tu et al., 1991 and 1992; Flower et al., 1992), many  
395 incompatible trace element ratios of Semporna lavas also resemble OIB. Mafic Semporna  
396 lavas possess LILE/LILE, HSFE/HFSE and LILE/HFSE ratios similar to OIB, although  
397 concentrations of Sr are less enriched than other LILE. These traits are also shared by  
398 central and northern Sulu Arc lavas (Fig. 9a). The marked decrease of Sr with decreasing  
399 MgO in the Semporna suite suggests that this may be a product of crustal contamination  
400 such that SBK13 underestimates the true Sr content of the Semporna source, relative to  
401 other elements (Fig. 3c). Similarly, the decrease of  $\text{P}_2\text{O}_5$  with MgO in most of the sub-groups  
402 (Fig. 2) means that the P content of the source may also be underestimated (Fig. 9a). All  
403 Sulu Arc lavas, however, have low LREE/HFSE and LREE/LILE, with respect to OIB. This is  
404 not a product of crustal contamination since the depletion of LREE relative to HFSE  
405 becomes less, not more, pronounced as MgO decreases (Fig. 5b). Indeed,  $(\text{Nb}/\text{La})_n$  is  
406 greater than 2 in the most mafic samples, suggesting a significant depletion of LREE with  
407 respect to the OIB source and to depleted mantle. Identifying a mechanism for producing  
408 LREE depletion of parental magma in the Sulu Arc would reconcile the differences between  
409 this and the South China Sea intra-plate magmatism (Fig. 9a) to a common OIB-like source.

410 The low LREE/HFSE and LREE/LILE ratios of mafic Sulu Arc magmatism might be  
411 produced in three ways: (1) through selective enrichment of LILE and HFSE in a source that  
412 was originally depleted in all incompatible elements, (2) during partial melting of an OIB

413 source in the presence of a phase that retains LREE, or (3) through element fractionation as  
414 melt migrates through the mantle.

415 The source of Plio-Pleistocene Sulu Arc magmatism cannot be produced through  
416 enrichment of depleted mantle by a slab-derived fluid. This process would generate the  
417 marked HFSE depletions typical of subduction-related magmatism. Partial melts of  
418 subducted slabs are also precluded as an enriching agent on the basis of isotopic ratios  
419 (Section 5.3.1). Metasomatism of depleted mantle by small degree partial melts of the upper  
420 mantle can fractionate LILE, HFSE and REE relative to one another, with or without  
421 producing modal metasomatic phases (Bodinier et al., 1996; Pilet et al., 2004). This  
422 explanation might be feasible for the Sulu Arc alone but is more difficult to sustain in view of  
423 the many other similarities between Sulu Arc and South China Sea lavas (Fig. 7 and 9b). A  
424 distinct LREE-enrichment event might have affected the South China Sea mantle,  
425 independent of an LILE- and HFSE-enrichment affecting the Sulu Arc and South China Sea.  
426 This, however, would require an entirely complementary relationship in the chemical budgets  
427 of the two metasomatic agents such that their summed effect produced a South China Sea  
428 mantle source with OIB-like chemistry. We consider this highly unlikely. Therefore, we  
429 conclude that enrichment of depleted mantle, alone, cannot have produced the similar  
430 sources of the Sulu Arc and South China Sea magmatic suites.

431 A wide range of minor, metasomatic phases might be present in the source of OIB-like  
432 magmas that could fractionate trace elements, particularly at low degree of partial melting. It  
433 is not possible to constrain possible roles for all of these but several obvious possibilities can  
434 be eliminated. Phlogopite and kaersutite can produce significant fractionation of HFSE from  
435 LREE but this should be in the opposite sense to that required to generate high Nb/La in  
436 Sulu i.e. Nb would be retained in the source relative to LREE yielding low-Nb/La melt  
437 (Schmidt et al., 1999; Tiepolo et al., 2000). This partitioning has strong compositional-  
438 dependence in amphibole but the ratio of partition coefficients approaches unity as the host  
439 rock Mg# approaches mantle values and does not reverse (Tiepolo et al., 2000). Apatite  
440 would preferentially retain LREE as well as P, which is mildly depleted in the Semporna  
441 rocks, but would also be expected to generate a significant negative Th anomaly (Chazot et  
442 al., 1996), which is not observed (Fig. 9a). Although other minor phases may be able to  
443 partition elements in a suitable way, the absolute concentration of incompatible elements in  
444 the different magmatic suites is inconsistent with derivation by variable degrees of partial  
445 melting of similar sources. The imprint of distinctive minor phases should be more apparent

446 in low degree partial melts. With increasing degrees of partial melting the residue would  
447 evolve towards a simpler assemblage with partition coefficients resembling those typical of  
448 the upper mantle and yield magma with lower concentrations of all incompatible elements.  
449 However, HFSE/LREE fractionation, with respect to OIB, is absent in South China Sea  
450 lavas, which possess the high incompatible element concentrations expected of lower  
451 degrees of partial melting (Fig. 9a). High Nb/La is found in the Sulu Arc lavas that have lower  
452 concentrations of incompatible elements. Therefore, we consider it unlikely that the LREE  
453 depletion of Sulu Arc magmatism results from low-degree partial melting of OIB source  
454 mantle in the presence of a residual phase with high  $D_{LREE}$ .

455 Interaction between melt and mantle peridotite may seem an unlikely process to produce the  
456 observed fractionation of LREE from LILE and HFSE, given that element partitioning should  
457 be governed by similar distribution coefficients to those operating during partial melting  
458 (Navon and Stolper, 1987). Indeed, experimental studies have concluded that reaction with  
459 peridotite will decrease the concentrations of HFSE in melt, relative to other incompatible  
460 elements (Kelemen et al., 1990 and 1993). Despite this, empirical evidence suggests that  
461 REE can be fractionated from other incompatible elements as basaltic melt interacts with  
462 upper mantle. Refertilization of depleted mantle by basaltic magma has been proposed as  
463 the origin of layered websterites ('group C' pyroxenites) from the "asthenospherised" part of  
464 the Ronda massif in Spain (Lenoir et al., 2001; Bodinier et al., 2008). Both the websterite  
465 layers and their host peridotites in the sub-lithospheric domain show strong enrichment of  
466 LREE relative to HFSE and LILE, while LILE/HFSE ratios display much less fractionation  
467 (Bodinier et al., 2008). A complementary (melt) product, with high HFSE/LREE and  
468 LILE/LREE ratios, is not observed at Ronda but the massif provides evidence that melt-  
469 mantle interaction can modify LREE concentrations of magma relative to elements with  
470 similar distribution coefficients.

### 471 *5.3.3 A model for intra-plate magmatism in the Sulu Arc and South China Sea*

472 Intra-plate magmatism in the Sulu Arc and South China Sea was derived from an OIB-like  
473 source. South China Sea magmatism occurred where the lithosphere was experiencing, or  
474 had recently experienced, mechanical thinning. Hainan, where most lava was erupted into  
475 the Lei-Qiong graben, was extended by pull-apart tectonics on the northern margin of the  
476 South China Sea (Tu et al., 1991; Flower et al., 1992). Reed Bank lies on edge of a  
477 presumed continental fragment on the conjugate, southern, extended margin of the inactive  
478 South China Sea. The mid- to late-Neogene Scarborough Seamounts were generated close



479 to the South China Sea spreading axis, which had become extinct 5-10 million years  
480 previously (Fig. 1a; Tu et al., 1992). The continental margin settings of Hainan and Reed  
481 Bank could be consistent with sources in the mantle lithosphere. However, the South China  
482 Sea lithosphere was very young when intra-plate magmatism occurred. Therefore, even if  
483 the source of this suite was hosted in the lithosphere it can have resided there for only a  
484 very short period since accretion/addition from the convecting mantle. In view of the  
485 widespread distribution of magmatic suites with similar trace element and isotopic chemistry  
486 (Fig. 9), we conclude that the source of intra-plate magmatism in the South China Sea and  
487 its extended margins was an OIB-like component in the upper mantle that melted as it  
488 upwelled beneath recently thinned lithosphere. As well as reducing the thickness of  
489 lithospheric mantle, preceding extension could provide large, lithosphere-penetrating  
490 structures that would facilitate transport of melt towards the surface and, thus, reduce the  
491 opportunity for interaction with the lithospheric mantle or crust.

492 Plio-Pleistocene magmatism in the Sulu Arc was extracted from similar enriched mantle,  
493 also during upwelling beneath thinned lithosphere. Upwelling might have occurred beneath  
494 localised sites of extension, but the Sulu Arc lithosphere would have experienced substantial  
495 thinning during Miocene subduction in the Sulu Arc (Andrews and Sleep, 1974; Hamilton,  
496 1995; Macpherson and Hall, 2002; Arcay et al., 2006). Such thinning is probably less reliant  
497 on mechanical deformation than is the case for extended margins. Instead, rheological  
498 changes result in (i) convective erosion, or corner flow, removing mass from the base of arc  
499 lithosphere (Hamilton, 1995; Billen and Gurnis, 2001; Arcay et al., 2006; Macpherson, 2008)  
500 and/or (ii) gravitational instabilities removing dense material from throughout the thickness of  
501 arc lithosphere (Rudnick, 1995). Lithospheric thin-spots produced by Miocene subduction  
502 would provide sites where enriched mantle could upwell and produce small volumes of intra-  
503 plate magmatism along the axis of the former arc front. In contrast to locations in and around  
504 the South China Sea, large, lithosphere-penetrating, extensional structures would be rare in  
505 the Sulu environment, increasing the probability that melt would interact with lithospheric  
506 mantle during transport from the subjacent asthenosphere.

507 Semporna lies at the end of the northeast-southwest trending Sulu Arc (Fig. 1a). Further  
508 southwest are several other Plio-Pleistocene to Recent low-volume volcanic fields that cap  
509 the topography of central Borneo at Hose Mountains, Kelian, Metalung, Nuit and Usun Apau  
510 (Fig. 1a). There are very few studies of these occurrences but data from Chiang (2002) show  
511 that basalt from Kelian possesses c. 20ppm Nb and there is a distinct decrease in Nb/K with

512 MgO, similar to that seen in the Sulu Arc (Fig. 5a). Therefore, the same enrichment  
513 responsible for Sulu Arc and South China Sea magmatism may be present in upper mantle  
514 beneath Borneo and has encountered thin spots in the lithosphere that have allowed it to  
515 upwell and melt. To the north of Borneo young, high-Nb magmatism from northern Palawan  
516 (Fig. 1a) may also share the same source (Arcilla et al., 2003).

#### 517 *5.4 Implications for high-Nb magmatism in active arcs*

518 Our findings indicate that high-Nb basaltic magmatism in the Semporna peninsula, and  
519 elsewhere in the Sulu Arc, does not require a contribution from subducted crust. If this  
520 conclusion is also valid in arcs where high-Nb magmatism is contemporaneous with typical  
521 arc magmatism then parts of the mantle wedge must escape significant metasomatism by  
522 material derived from the subducted slab. In particular, the original finding of Reagan and  
523 Gill (1989); that high-Nb basalts and calc-alkaline magma were erupted from the same  
524 centre, implies that slab-fluxed and un-fluxed mantle may be present within the source  
525 volume of a single volcano.

526 Intra-plate or OIB-type mantle has been advocated as the prevalent mantle wedge  
527 component in some volcanic arcs (e.g. Mexico, Gómez-Tuena et al., 2007). In view of the  
528 low recycled flux inferred for Semporna mantle it is tempting to regard the source of Plio-  
529 Pleistocene magmatism as representative of the bulk mantle beneath the Sulu Arc.  
530 However, the relatively low volumes of Semporna Plio-Pleistocene magmatism indicate a  
531 finite source that was rapidly exhausted after melting commenced. This conclusion is  
532 reinforced by the other sites in Borneo and the South China Sea, where OIB-like magma  
533 also occurs in low volumes. Such enriched domains may be relatively common in the upper  
534 mantle beneath much of SE Asia, but their signature would be swamped when conditions  
535 allow partial melting of the more refractory mantle in which the enrichments are hosted. This  
536 is analogous to the recognition of melt derived from enriched mantle on the margins of active  
537 rift systems. Such domains may also be present beneath the rift but their signature is  
538 overwhelmed where partial melting becomes more extensive close to rift axes (e.g. Iceland,  
539 Fitton et al., 2003).

#### 540 *5.5 The high-Nb basalt – adakite association*

541 The frequent association of high-Nb basalt with adakitic magmatism led Defant et al. (1992)  
542 to postulate a genetic link, in which adakitic magma metasomatised the mantle to produce

543 the high-Nb source. Defant and Drummond (1990) regarded adakites as direct samples of  
544 magma generated by partial melting of subducted basaltic crust but an increasing number of  
545 studies have questioned this model (Garrison and Davidson, 2003; Prouteau and Scaillet,  
546 2003; Chiaradia et al., 2004; Macpherson et al., 2006; Eiler et al., 2007; Rodriguez et al.,  
547 2007). The sources of high-Nb basalt in Semporna, and related SE Asian sites, cannot have  
548 been produced by metasomatism of depleted mantle by slab melt (Section 5.3.1).  
549 Furthermore, although adakitic rocks have been found in the northern Sulu Arc (Sajona et al.  
550 1996; Castillo et al., 2007) there is no evidence for adakitic magmatism in the Semporna  
551 peninsula. Similarly, adakitic magmatism has not been documented in association with the  
552 high-Nb basalt suites of the South China Sea and its margins. Therefore, Semporna and the  
553 South China Sea weaken the case for a petrogenetic link where these two distinctive  
554 “flavours” of magmatism occur in a single subduction zone.

555 Despite this assertion, the fact remains that several margins have produced both adakitic  
556 and high-Nb magmatism (Defant et al., 1992). Macpherson et al. (2006) used an example  
557 from the East Philippine Arc to show that adakitic magmatism can occur where hydrous arc  
558 basalt, produced by fluid-fluxed melting of the mantle wedge, ponds at relatively deep levels  
559 and crystallises garnet ( $\pm$  amphibole). Adakitic magma was produced when this crystal  
560 assemblage was removed from hydrous basaltic magma or when the resulting cumulate  
561 rocks experienced partial melting. The East Philippine setting was conducive to these  
562 processes because the plate margin was young and so the arc lithosphere had experienced  
563 limited thinning. The deeper parts of this thick arc lithosphere acted as a barrier to melt  
564 transport promoting crystallisation of basalt at depth. Meanwhile, the shallow portions had  
565 yet to develop substantial magma plumbing systems, therefore, geochemical evidence of  
566 deep differentiation was not overprinted by subsequent differentiation when the adakitic melt  
567 was emplaced (Macpherson, 2008).

568 The potential for an active arc to generate high-Nb magmatism depends on the presence of  
569 a suitably enriched source in the mantle wedge. But, like adakitic magma, the distinctive  
570 geochemistry of high-Nb basalt is most likely to be preserved where it is not overprinted by  
571 interaction with large volumes of melt derived from slab-modified mantle wedge and/or by  
572 differentiation in the shallow crust. We propose that the occurrence of adakitic and high-Nb  
573 magmatism together in an arc does not reflect a genetic link between their sources. Instead,  
574 we postulate that there is a significant increase in the probability that magmatism will retain  
575 distinctive geochemical signatures derived at depth e.g. either by deep differentiation

576 (adakitic magmatism) or inherited from a distinctive mantle source (high-Nb magmatism),  
577 during transport through arc lithosphere that receives a low flux of melt from slab-modified  
578 mantle and/or hosts poorly developed magma plumbing in the shallow crust.

## 579 **6. Conclusions**

580 1. Plio-Pleistocene basalts and basaltic andesites from the Semporna peninsula of the  
581 southern Sulu Arc contain higher concentrations of Nb than typical arc magmatism. The  
582 most mafic lavas have negligible Nb depletions relative to elements with similar compatibility.  
583 Depletion of Nb, and several other incompatible elements, occurred during differentiation  
584 from basalt to basaltic andesite. This was accompanied by striking changes in isotopic ratios  
585 that indicate interaction with the crust. The isotopic characteristics of the contaminant  
586 indicate an ancient, possibly Archean, component is present in the Sabah crust.

587 2. The primitive Semporna lavas closely resemble high-Nb and Nb-enriched basalts from the  
588 central (Sulu Islands) and northern (Zamboanga) segments of the Sulu Arc. Isotopic ratios  
589 preclude a role for metasomatism of Sulu Arc mantle by melt derived from subducted Sulu  
590 Sea or Celebes Sea oceanic crust. Mafic Sulu Arc lavas possess incompatible trace element  
591 ratios that resemble ocean island basalt but are depleted in light rare earth elements. Sulu  
592 Arc basalts also resemble mafic magmatism at several sites in and around the South China  
593 Sea, which differ only in lacking light rare earth element depletion. This similarity and the  
594 range of localities indicates a common source present in the convecting upper mantle. This  
595 magmatic province may also extend southwest into central Borneo.

596 3. The Sulu Arc runs from the Zamboanga peninsula through the Sulu Islands to the  
597 Semporna peninsula, yet there is little other geological or geophysical evidence to support  
598 active subduction beneath this structure. Plio-Pleistocene magmatism resulted from  
599 upwelling of OIB-like domains in the upper mantle into lithospheric thin spots that were  
600 produced during Miocene subduction.

601 4. Light rare earth element depletion of Sulu magmatism cannot be attributed to crustal  
602 contamination and probably occurred when basaltic melt interacted with mantle peridotite  
603 during transport through the Sulu Arc lithosphere. South China Sea magmatism may have  
604 escaped this process due to transport along extensional structures in oceanic lithosphere  
605 and stretched continental margins.

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## 864 **Figure Captions**

865 Figure 1. (a) Map showing location of Borneo and other sites discussed with plate  
866 boundaries (thick solid lines) and major faults (dashed lines). Black box shows location of  
867 (b). Geographic features are noted in capitals. Locations of magmatic suites plotted in  
868 Figures 4, 5 and 9 are listed in italics. Abbreviations; H – Hose Mountains, K – Kelian, M –  
869 Metalung, N – Nuit, NP - North Palawan, U - Usun Apau. (b) Semporna peninsula in  
870 southeastern Sabah showing the distribution of Mio-Pliocene and Plio-Pleistocene  
871 magmatism after Kirk (1962), Haile et al. (1965), Leong (1974), Lim (1981), Lee (1988),  
872 Bellon and Rangin (1991) and Hutchison et al. (2000).

873 Figure 2. Plots of selected major elements versus MgO for Plio-Pleistocene lavas from  
874 Tawau (PP1 – PP4) and Mostyn.

875 Figure 3. Plots of selected trace elements versus MgO for Plio-Pleistocene lavas from  
876 Tawau (PP1 – PP4) and Mostyn.

877 Figure 4. N-MORB normalised multi-elements plots for Plio-Pleistocene lavas from Tawau  
878 and Mostyn. All normalisation factors from Sun and McDonough (1989).

879 Figure 5. Plots of (a) Nb/K, and (b) Nb/La, normalised to N-MORB, versus MgO for Plio-  
880 Pleistocene lavas from Tawau and Mostyn. Data for Kelian, central Borneo, from Chiang  
881 (2002).

882 Figure 6. Plots of (a)  $^{87}\text{Sr}/^{86}\text{Sr}$ , (b)  $^{143}\text{Nd}/^{144}\text{Nd}$  and (c)  $^{206}\text{Pb}/^{204}\text{Pb}$  versus MgO for Plio-  
883 Pleistocene lavas from Tawau and Mostyn. Mio-Pliocene arc data from Tawau from Chiang  
884 (2002).

885 Figure 7. (a)  $^{143}\text{Nd}/^{144}\text{Nd}$  versus  $^{87}\text{Sr}/^{86}\text{Sr}$ , (b)  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$ , (c)  $^{208}\text{Pb}/^{204}\text{Pb}$   
886 versus  $^{206}\text{Pb}/^{204}\text{Pb}$ , and (d)  $^{143}\text{Nd}/^{144}\text{Nd}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  for Plio-Pleistocene lavas from  
887 Tawau and Mostyn. Comparison data shown for Indian MORB (GERM:  
888 <http://earthref.org/GERM/>); northern and central Sulu Arc (Castillo et al., 2007); Sulu and  
889 Celebes seafloor basalts (Spadea et al., 1996), Scarborough Seamounts and Reed Banks  
890 (Tu et al., 1992) and Hainan Island (Tu et al., 1991). Northern Hemisphere Reference Line in  
891 (b) and (c) from Hart (1984).

892 Figure 8. Comparison of  $^{87}\text{Sr}/^{86}\text{Sr}$  versus  $^{143}\text{Nd}/^{144}\text{Nd}$  of Plio-Pleistocene lavas from  
893 Semporna with assimilation with fractional crystallisation (AFC) models (DePaolo, 1981). In  
894 each case the uncontaminated melt has isotopic ratios and trace element concentrations of  
895 basalt SBK13. In all AFC models  $r = 0.15$ , except EIS 2 in which  $r = 0.85$ . Tick marks  
896 represent reduction of the fraction of melt remaining by 0.1, except for EIS 2 where ticks are  
897 shown for 0.99, 0.98 and 0.95 of original melt. Partition coefficients are 1.5 for Sr and 0.1 for  
898 Nd (GERM: <http://earthref.org/GERM/>). The contaminants for each model are: LC 1,  $^{87}\text{Sr}/^{86}\text{Sr}$   
899 = 0.709259,  $^{143}\text{Nd}/^{144}\text{Nd} = 0.512304$ , Sr = 252ppm and Nd = 23.2ppm (Capoas granite,  
900 Palawan; Encarnación and Mukasa, 1997); EIS 1 and 2,  $^{87}\text{Sr}/^{86}\text{Sr} = 0.739404$ ,  $^{143}\text{Nd}/^{144}\text{Nd} =$   
901 0.511984, Sr = 114ppm and Nd = 38.1ppm (East Indonesian Sediment; Vroon et al., 1995);  
902 AC 1,  $^{87}\text{Sr}/^{86}\text{Sr} = 0.72460$ ,  $^{143}\text{Nd}/^{144}\text{Nd} = 0.51025$ , Sr = 400ppm and Nd = 43.1ppm  
903 (Beartooth Mountains, USA; Wooden and Mueller, 1988); AC2,  $^{87}\text{Sr}/^{86}\text{Sr} = 0.70890$ ,  
904  $^{143}\text{Nd}/^{144}\text{Nd} = 0.51041$ , Sr = 573ppm and Nd = 28.8ppm (Archean migmatite from Lofoten-  
905 Verterålen, Norway; Jacobsen and Wasserburg, 1978).

906 Figure 9. (a) OIB normalised multi-elements plots for mafic lavas from Semporna, northern  
907 Sulu Arc, Reed Bank and Hainan Island. (b)  $^{143}\text{Nd}/^{144}\text{Nd}$  versus Zr/Nb for high-Nb basalts  
908 from Semporna peninsula, Sulu Arc and South China Sea sites. Data sources as in Fig. 7.

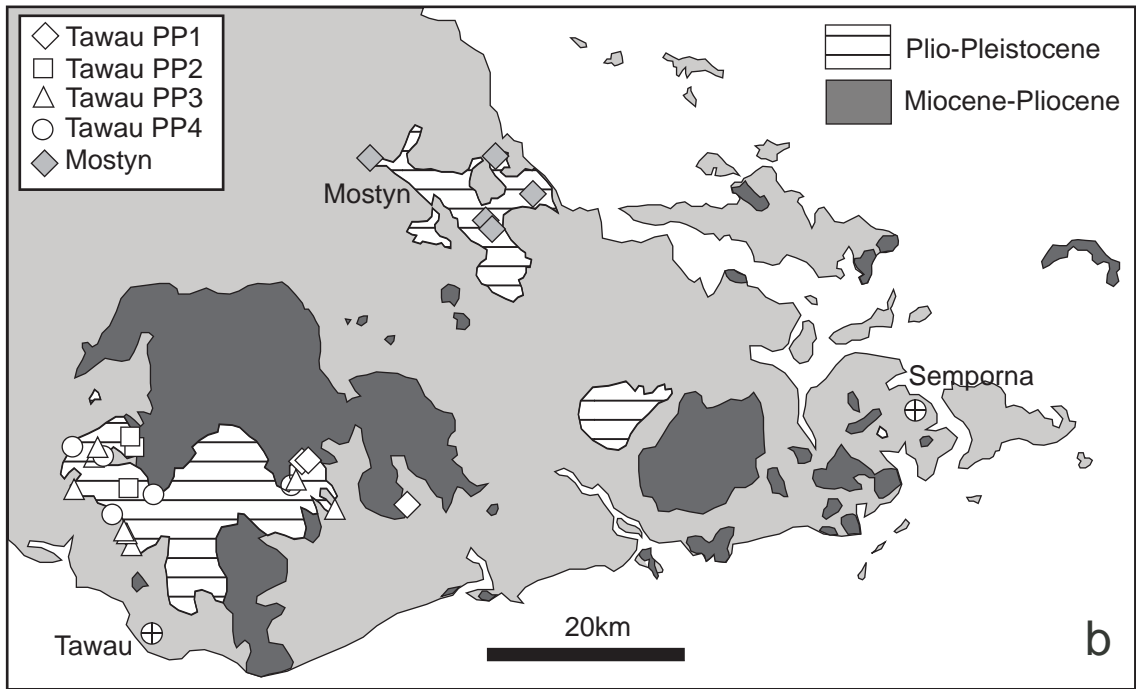
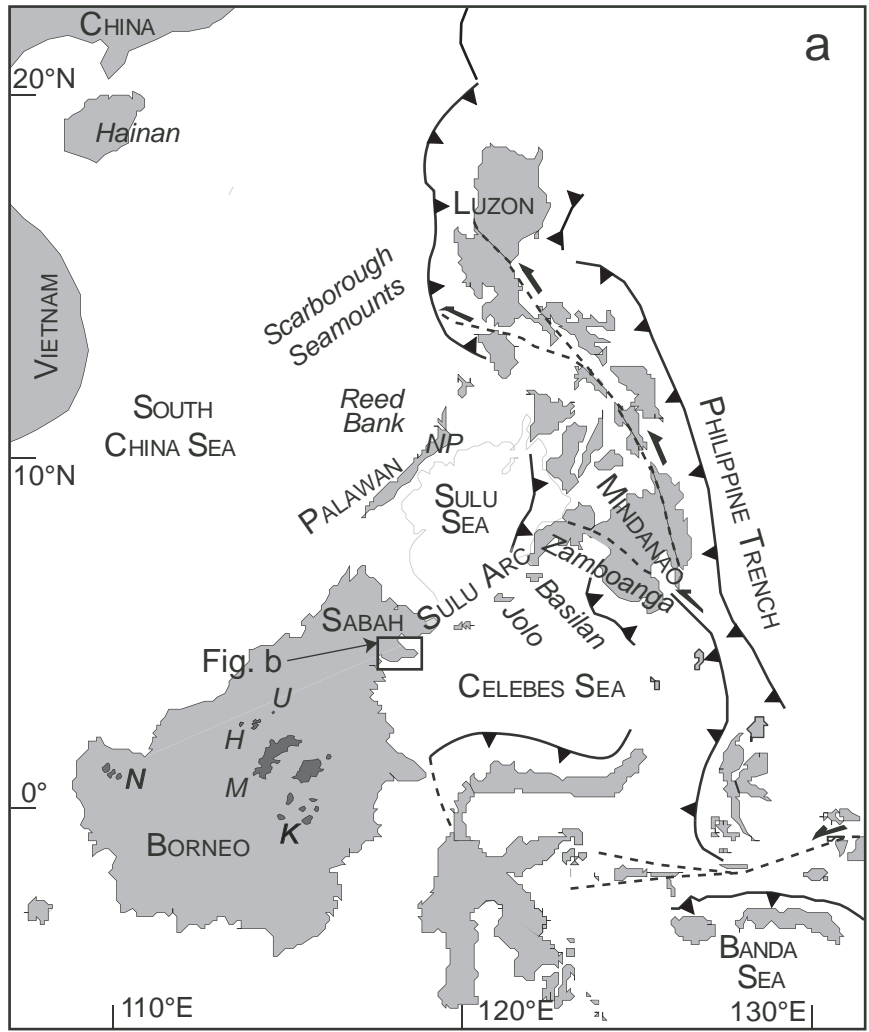


Figure 1

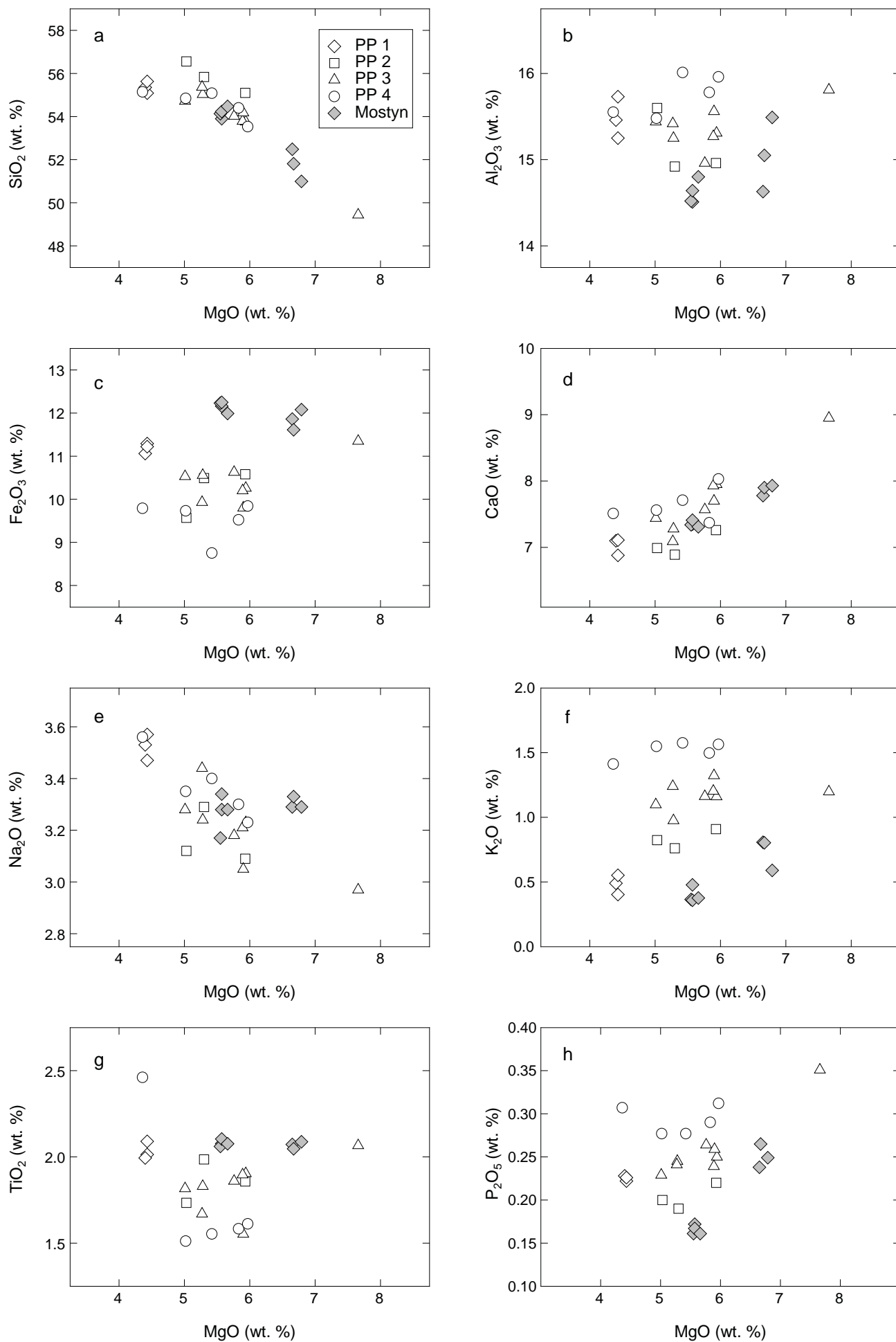


Figure 2

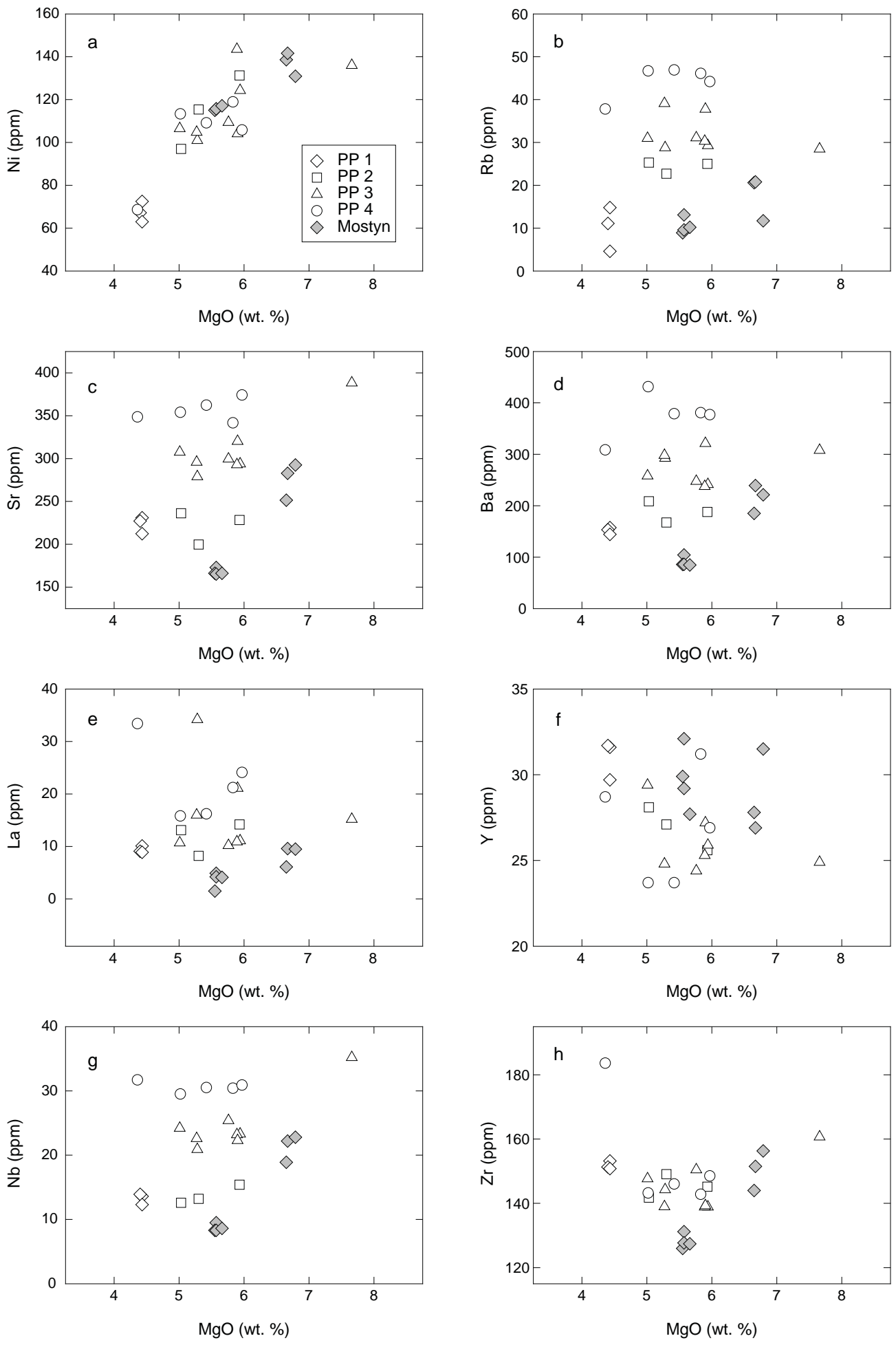


Figure 3



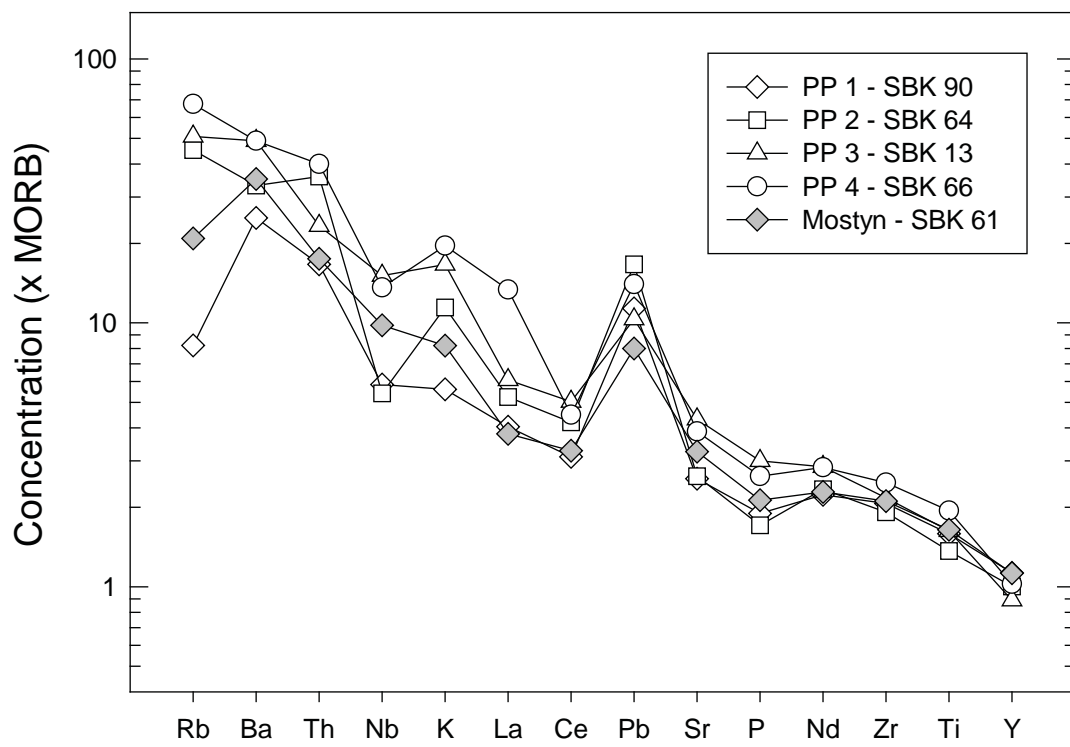


Figure 4

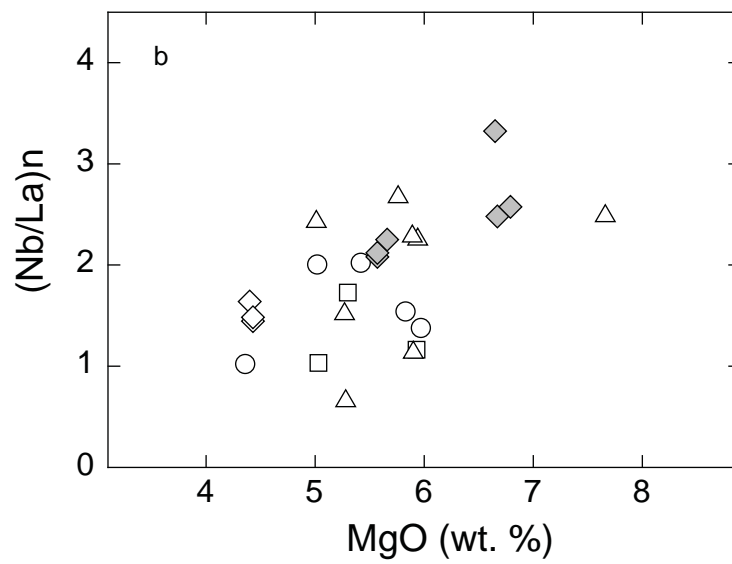
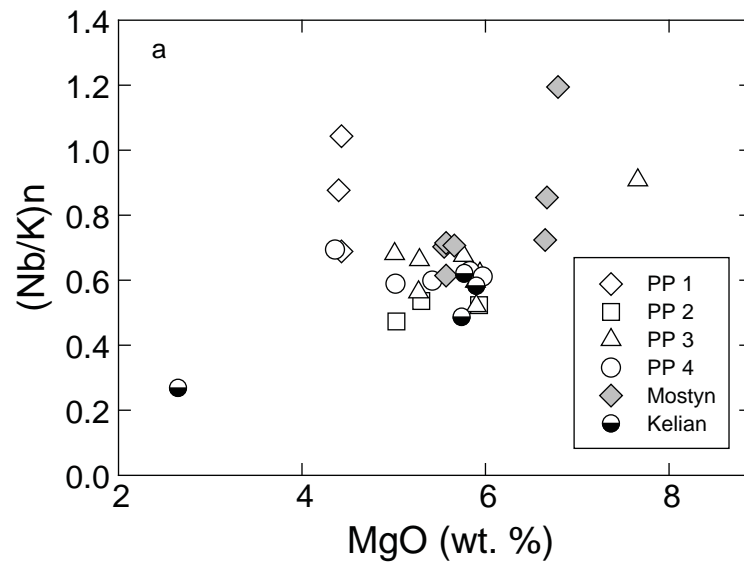


Figure 5

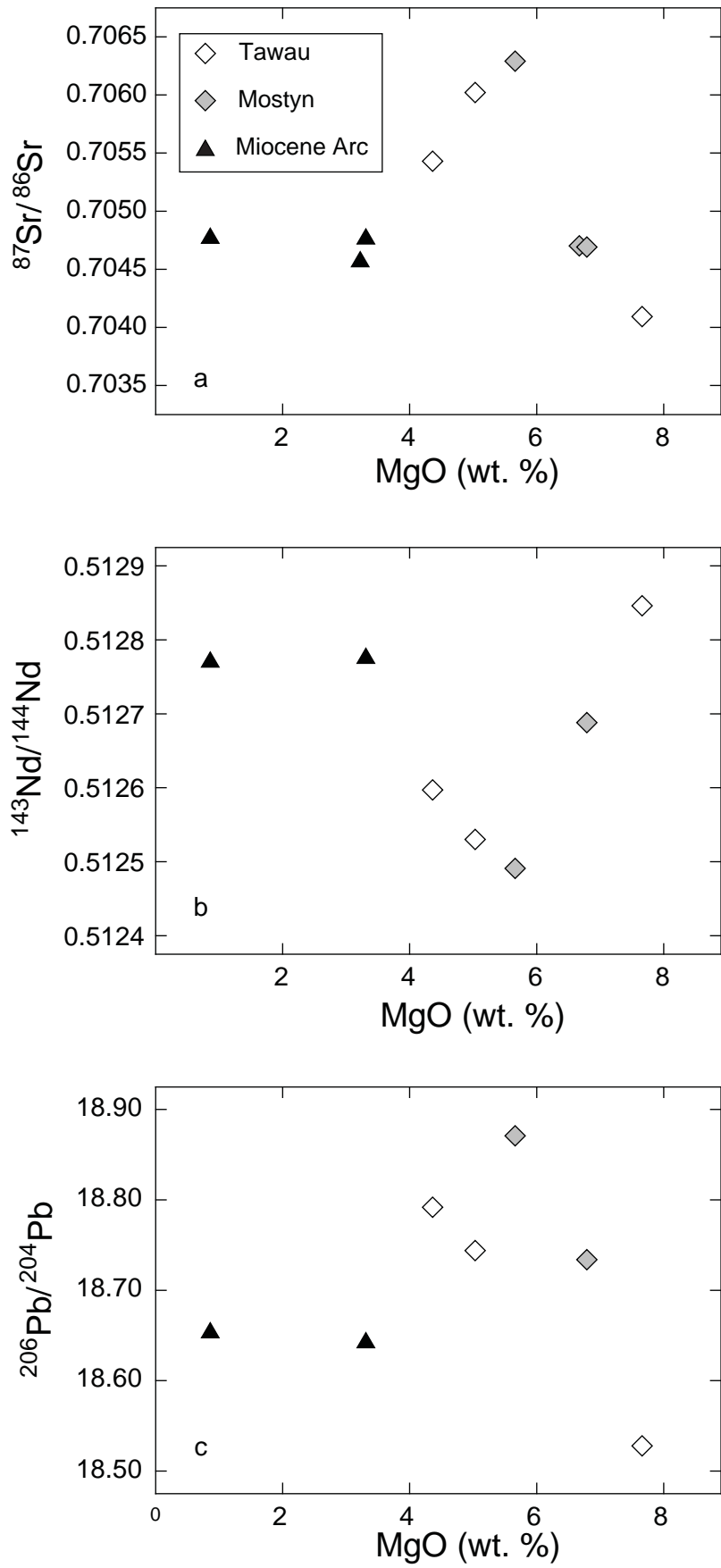


Figure 6

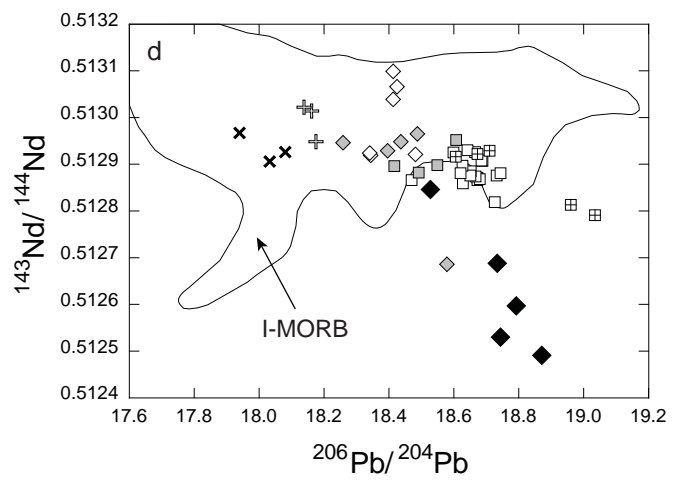
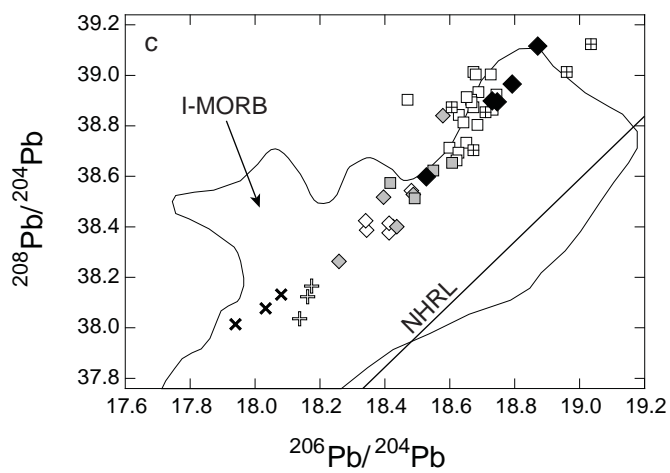
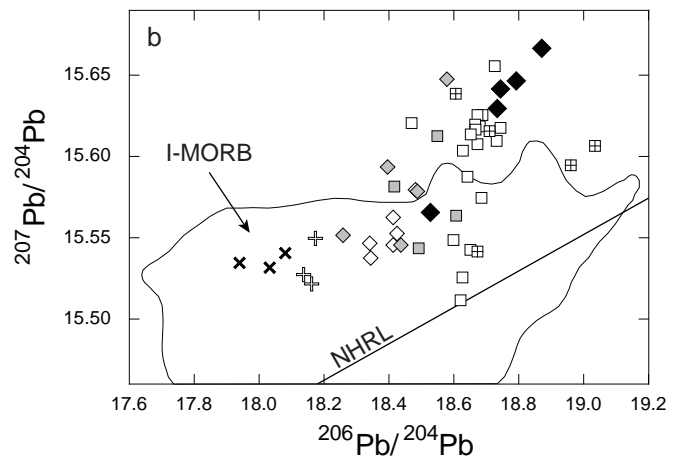
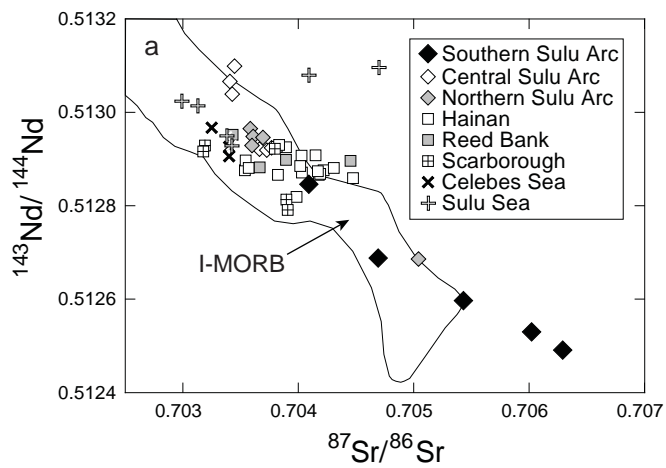


Figure 7

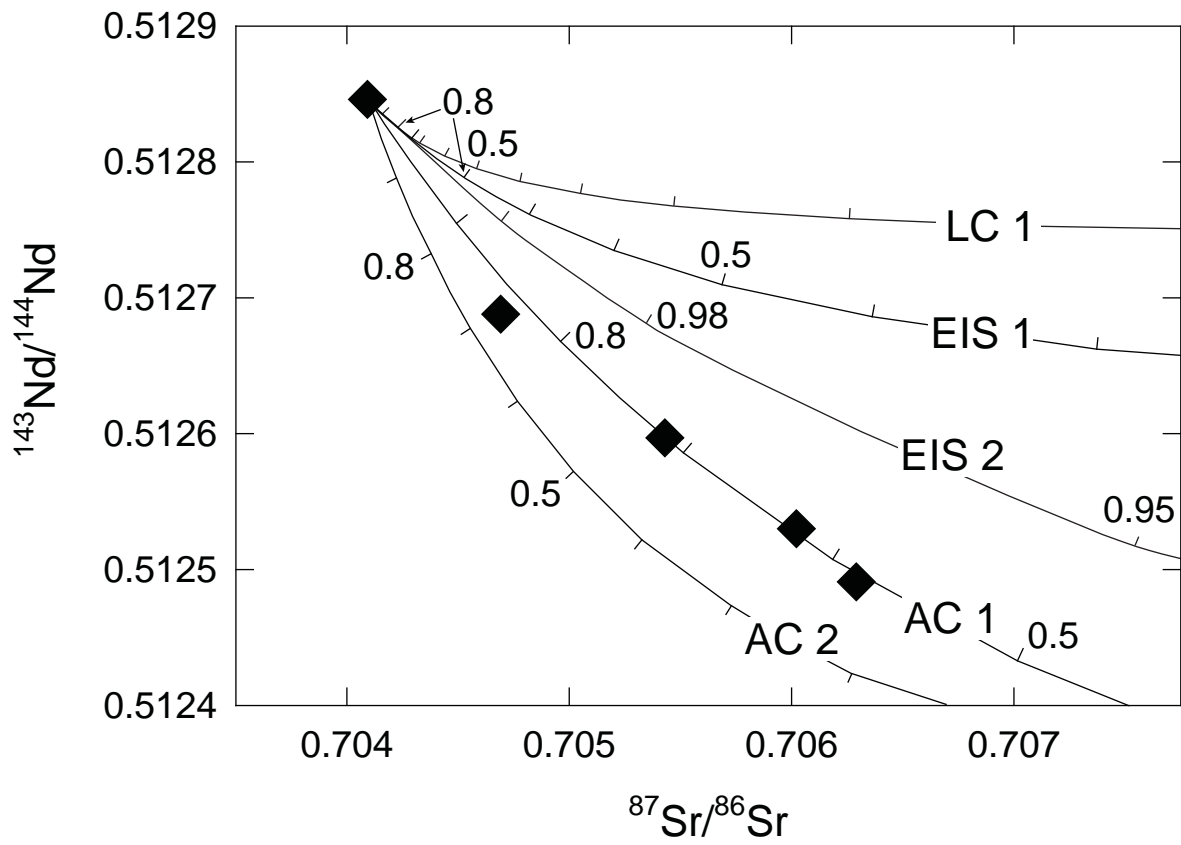


Figure 8

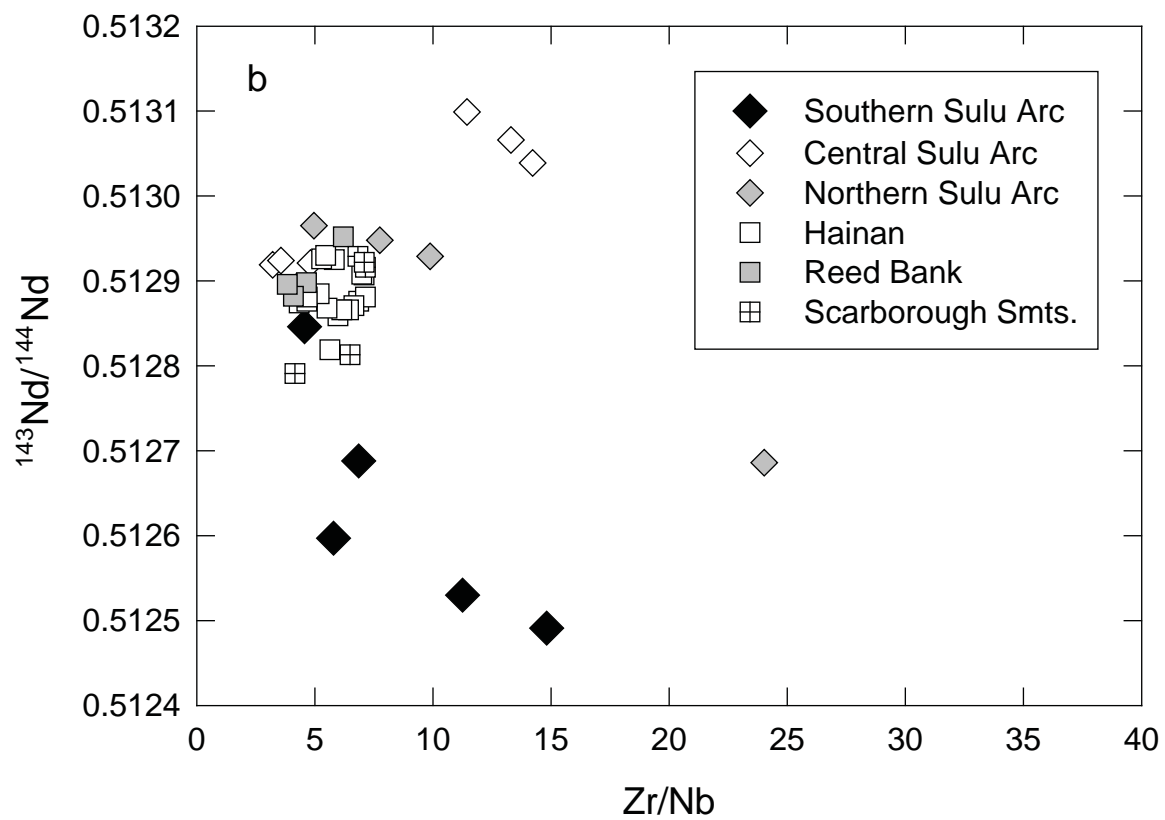
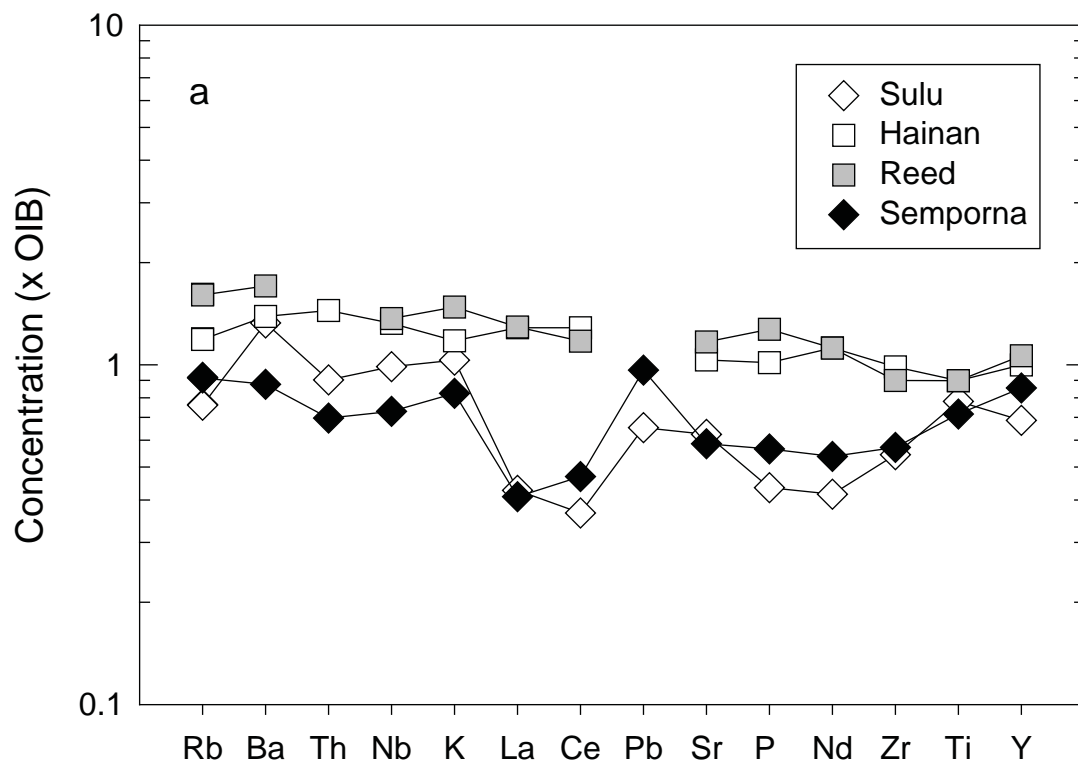


Figure 9

Table 1. Major and trace element concentrations in Plio-Pleistocene lavas from the Semporna Peninsula.

	Tawau SBK3	Tawau SBK5	Tawau SBK6	Tawau SBK7	Tawau SBK13	Tawau SBK64	Tawau SBK65	Tawau SBK66	Tawau SBK67	Tawau SBK68	Tawau SBK69	Tawau SBK70	Tawau SBK71
SiO <sub>2</sub> (wt.%, ± 0.16)	55.08	54.84	54.02	54.72	49.44	56.56	55.84	55.14	55.03	55.37	53.76	53.79	55.10
TiO <sub>2</sub> (wt.%, ± 0.010)	1.55	1.51	1.86	1.82	2.07	1.73	1.99	2.46	1.83	1.67	1.90	1.90	1.86
Al <sub>2</sub> O <sub>3</sub> (wt.%, ± 0.07)	16.01	15.48	14.96	15.44	15.81	15.60	14.92	15.55	15.25	15.42	15.31	15.27	14.96
Fe <sub>2</sub> O <sub>3</sub> (wt.%, ± 0.10)	8.75	9.73	10.63	10.53	11.35	9.57	10.49	9.79	10.56	9.93	10.26	10.20	10.58
MgO (wt.%, ± 0.07)	5.42	5.02	5.76	5.01	7.66	5.03	5.30	4.36	5.28	5.27	5.94	5.89	5.93
MnO (wt.%, ± 0.006)	0.14	0.23	0.15	0.14	0.17	0.14	0.14	0.14	0.17	0.14	0.15	0.14	0.14
CaO (wt.%, ± 0.03)	7.71	7.56	7.57	7.44	8.95	6.99	6.89	7.51	7.28	7.09	7.95	7.93	7.26
Na <sub>2</sub> O (wt.%, ± 0.13)	3.40	3.35	3.18	3.28	2.97	3.12	3.29	3.56	3.24	3.44	3.23	3.21	3.09
K <sub>2</sub> O (wt.%, ± 0.010)	1.58	1.55	1.16	1.10	1.20	0.82	0.76	1.41	0.98	1.24	1.16	1.20	0.91
P <sub>2</sub> O <sub>5</sub> (wt.%, ± 0.007)	0.28	0.28	0.26	0.23	0.35	0.20	0.19	0.31	0.25	0.24	0.25	0.24	0.22
Total	99.91	99.55	99.56	99.70	99.97	99.76	99.80	100.24	99.86	99.81	99.91	99.77	100.04
LOI	-0.44	0.20	-0.39	0.20	0.84	0.11	-0.49	-0.20	0.01	-0.43	-0.21	-0.47	-0.10
Mg#	55.1	50.6	51.8	48.5	57.2	51.0	50.0	46.9	49.8	51.2	53.4	53.3	52.6
Ni (ppm, ± 1.0)	109.1	113.3	109.4	106.5	136.0	97.0	115.4	68.6	101.0	104.8	124.3	143.5	131.3
Cr (± 1.0)	168.8	160.7	159.4	154.5	231.2	164.8	188.7	73.9	151.6	153.3	202.9	203.1	218.9
V (± 1.0)	149.9	144.6	151.6	156.9	240.1	160.0	159.8	171.9	141.5	140.3	165.2	165.8	165.2
Sc (± 0.6)	19.1	19.3	19.8	20.2	25.5	19.6	19.6	20.8	19.4	19.1	21.7	22.5	22.1
Cu (± 1.0)	44.3	42.3	46.5	35.8	61.3	46.1	46.2	45.7	47.4	46.8	57.1	56.9	51.2
Zn (± 0.8)	94.2	90.8	106.7	110.6	106.2	104.7	116.0	110.2	104.7	99.6	112.4	110.8	109.6
Cl (± 50)	409	170	63	97				206	163	189	204	236	114
Ga (± 0.7)	20.3	18.8	18.9	20.1	21.1	21.3	21.0	20.9	20.3	20.1	21.4	20.1	19.3
Ba (± 3)	379	432	248	258	308	209	167	308	293	298	242	238	188
Rb (± 0.4)	46.9	46.7	31.1	31.0	28.5	25.3	22.7	37.8	28.8	39.1	29.3	30.3	25.0
Sr (± 0.6)	362.3	353.9	299.7	307.6	388.4	236.2	199.8	348.6	279.0	296.0	293.9	293.0	228.3
Zr (± 0.6)	146.0	143.2	150.5	147.7	160.7	141.8	149.1	183.6	144.3	139.0	138.9	138.9	145.2
Nb (± 0.3)	30.5	29.5	25.4	24.2	35.2	12.6	13.2	31.7	20.9	22.6	23.3	23.2	15.4
Y (± 0.4)	23.7	23.7	24.4	29.4	24.9	28.1	27.1	28.7	65.8	24.8	25.9	25.3	25.6
La (± 1.0)	16.2	15.8	10.2	10.7	15.2	13.1	8.2	33.4	34.2	16.0	11.1	10.9	14.2
Ce (± 3.0)	33.6	35.4	25.8	26.5	37.7	31.5	24.5	33.6	33.5	26.3	26.5	25.5	29.6
Nd (± 0.7)	18.4	18.8	16.0	17.2	20.8	17.1	15.7	20.7	41.3	16.3	16.3	15.7	17.0
Pb (± 0.9)	4.5	5.9	2.6	4.8	3.1	5.0	4.4	4.2	3.1	4.1	2.9	3.3	5.3
Th (± 0.7)	4.4	4.0	2.7	3.4	2.8	4.3	4.4	4.8	2.4	3.8	3.7	2.3	5.9

Table 1 (cont.).

	Tawau SBK72	Tawau SBK90	Tawau SBK91	Tawau SBK92	Tawau SBK93	Tawau SBK94	Mostyn SBK30	Mostyn SBK31	Mostyn SBK60	Mostyn SBK61	Mostyn SBK62	Mostyn SBK63	Mostyn SA9802
SiO <sub>2</sub> (wt.%)	54.40	55.09	55.35	55.63	54.18	53.53	53.90	52.48	51.81	50.99	54.13	54.23	54.47
TiO <sub>2</sub>	1.58	2.01	1.99	2.09	1.55	1.61	2.08	2.07	2.05	2.09	2.06	2.10	2.08
Al <sub>2</sub> O <sub>3</sub>	15.78	15.73	15.46	15.25	15.56	15.96	14.51	14.63	15.05	15.49	14.52	14.64	14.80
Fe <sub>2</sub> O <sub>3</sub>	9.52	11.29	11.06	11.22	9.80	9.84	12.16	11.86	11.61	12.08	12.23	12.25	11.99
MgO	5.83	4.43	4.40	4.43	5.90	5.97	5.57	6.65	6.67	6.79	5.55	5.57	5.66
MnO	0.15	0.13	0.15	0.15	0.14	0.14	0.16	0.15	0.16	0.17	0.14	0.16	0.16
CaO	7.37	6.88	7.10	7.11	7.70	8.03	7.35	7.78	7.90	7.93	7.34	7.41	7.31
Na <sub>2</sub> O	3.30	3.47	3.53	3.57	3.05	3.23	3.34	3.29	3.33	3.29	3.17	3.28	3.28
K <sub>2</sub> O	1.50	0.40	0.49	0.55	1.32	1.56	0.48	0.81	0.80	0.59	0.37	0.36	0.38
P <sub>2</sub> O <sub>5</sub>	0.29	0.22	0.23	0.23	0.26	0.31	0.17	0.24	0.27	0.25	0.16	0.17	0.16
Total	99.71	99.66	99.77	100.22	99.46	100.18	99.73	99.97	99.65	99.66	99.67	100.17	100.29
LOI	-0.07	0.77	0.03	-0.53	0.07	0.04	-0.50	-0.70	-0.44	0.13	-0.44	-0.59	-0.34
Mg#	54.8	43.8	44.1	43.9	54.4	54.6	47.6	52.6	53.2	52.7	47.3	47.4	48.3
Ni (ppm)	118.9	72.5	67.2	63.0	104.1	105.8	115.6	138.6	141.6	130.9	115.0	115.7	117.1
Cr	178.9	120.3	122.5	114.1	177.3	193.5	186.2	214.9	219.3	218.2	192.3	190.1	191.9
V	148.2	154.3	153.1	147.9	160.1	165.2	164.8	163.0	167.1	173.3	164.5	164.6	164.9
Sc	19.9	21.1	20.1	19.9	21.1	21.6	21.2	21.2	22.3	23.5	22.1	23.6	22.6
Cu	47.3	39.8	40.8	45.1	44.2	44.6	47.1	51.2	51.2	47.5	45.3	45.7	48.6
Zn	88.4	119.5	120.6	114.0	98.4	93.2	123.2	115.1	115.1	112.9	122.1	122.0	116.2
Cl	181		7		150	215	94	165	134	115			
Ga	20.2	21.6	22.3	21.6	18.7	20.8	20.8	20.9	20.8	21.0	20.3	21.8	21.0
Ba	381	157	153	144	321	377	104	185	239	221	86	86	85
Rb	46.1	4.6	11.1	14.8	37.8	44.2	13.1	20.6	20.8	11.7	8.9	9.6	10.2
Sr	341.7	231.0	227.2	212.4	320.3	374.1	172.9	251.4	282.8	292.5	166.2	165.3	166.2
Zr	142.8	153.2	151.3	150.8	139.3	148.5	131.2	144.0	151.5	156.3	126.0	127.7	127.4
Nb	30.4	13.6	13.9	12.3	22.3	30.9	9.5	18.9	22.2	22.8	8.3	8.3	6.8
Y	31.2	31.6	31.7	29.7	27.2	26.9	32.1	27.8	26.9	31.5	29.9	29.2	27.7
La	21.2	10.1	9.1	8.9	21.1	24.1	4.9	6.1	9.6	9.5	1.5	4.2	4.1
Ce	35.0	23.3	23.1	22.6	36.7	41.6	15.1	19.3	24.7	24.6	13.6	11.2	14.5
Nd	20.6	16.3	17.5	15.5	20.9	24.4	12.6	14.5	15.5	16.7	11.7	11.1	12.3
Pb	4.8	3.4	4.1	2.4	5.7	5.7	2.8	1.8	2.4	2.4	2.6	2.7	2.0
Th	4.6	2.0	2.8	2.8	5.8	6.3	1.4	2.6	2.9	2.1	1.5	1.3	1.9



Table 2. Isotopic ratios of Plio-Pleistocene lavas from the Semporna Peninsula.

	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
SBK13	0.704092	0.512846	18.528	15.566	38.598
SBK64	0.706021	0.512530	18.744	15.642	38.895
SBK66	0.705430	0.512597	18.792	15.647	38.966
SBK60	0.704701				
SBK61	0.704691	0.512688	18.734	15.630	38.899
SA9802	0.706291	0.512491	18.871	15.667	39.116