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# Petrogenesis of plagiogranites in the Muslim Bagh Ophiolite, Pakistan

Cox, Daniel; Kerr, Andrew; Hastie, Alan; Kakarc, Ishaq

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**1** **Petrogenesis of plagiogranites in the Muslim Bagh Ophiolite,**

**2** **Pakistan: implications for the generation of Archean continental**

**3** **crust**

4

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7 Daniel Cox[a,b*], Andrew C. Kerr[b], Alan R. Hastie[a], M. Ishaq Kakar[c]

8

9 [a] School of Geography, Earth and Environmental Sciences, University of Birmingham,

10 Edgbaston, Birmingham, B15 2TT, U.K.

11 [b] School of Earth and Ocean Sciences, Cardiff University, Main Building, Park Place,

12 Cardiff, CF10 3AT, U.K.

13 [c] Centre of Excellence in Mineralogy, University of Balochistan, Quetta, Pakistan.

14

15 * Corresponding author. E-mail: DXC506@student.bham.ac.uk, School of Geography, Earth

16 and Environmental Sciences, University of Birmingham, Edgbaston, Birmingham, B15 2TT,

17 U.K.

18

19 Andrew C. Kerr: KerrA@cardiff.ac.uk; Alan R. Hastie: A.R.Hastie@bham.ac.uk;

20 M. I. Kakar: kakarmi.cemuob@gmail.com.

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23

**24  Abstract**

25  High-SiO$_2$ rocks referred to as oceanic plagiogranites are common within the crustal

26  sequences of ophiolites; however, their mode of petrogenesis is controversial with both late-

27  stage fractional crystallisation and partial melting models being proposed. Here, we present

28  new whole rock data from plagiogranitic dyke-like bodies and lenses from the lower and

29  middle sections of the sheeted dyke complex of the Cretaceous Muslim Bagh Ophiolite,

30  north-western Pakistan. The plagiogranites have similar geochemical signatures that are

31  inconsistent with them being the fractionation products of the mafic units of the Muslim

32  Bagh Ophiolite. However, the plagiogranites all display very low TiO$_2$ contents (<0.4 wt. %),

33  implying that they formed by partial melting of mafic rocks. Melt modelling of a crustal

34  gabbro from the Muslim Bagh Ophiolite shows that the trace element signature of the

35  plagiogranites can be replicated by 5 − 10% melting of a crustal hornblende gabbro, with

36  amphibole as a residual phase, resulting in a concave-up middle rare earth element pattern.

37  Compositional similarities between the Muslim Bagh Ophiolite plagiogranites and Archean

38  TTG (trondhjemite − tonalite − granodiorite) has implications for the generation of juvenile

39  Archean continental crust. As the Muslim Bagh Ophiolite was derived in a supra-subduction

40  zone, it is suggested that some Archean TTG may have been derived from melting of mafic

41  upper crust in early subduction-like settings. However, due to the small volume of Muslim

42  Bagh Ophiolite plagiogranites, it is inferred that they can be instructive on the petrogenesis of

43  some, but not all, of Archean TTG.

44

45  **Keywords**: Pakistan, Muslim Bagh, Ophiolite, Oceanic Plagiogranite, Partial Melting

46

47  **1. Introduction**

48    Within obducted Phanerozoic ophiolite sequences, suites of felsic rocks termed "oceanic

49    plagiogranites" (Coleman & Peterman, 1975; Le Maitre et al., 2002, p. 118) occur as small

50    volume (<10%) components (Coleman & Peterman, 1975; Koepke et al., 2007). The

51    petrogenesis of these plagiogranites is controversial, having been variously proposed to have

52    formed by the late-stage crystallisation of mafic melts (Coleman & Peterman, 1975), hydrous

53    partial melting (and assimilation) of mafic rocks (Gerlach et al., 1981; Amri et al., 1996;

54    Gillis & Coogan, 2002; France et al., 2009, 2010; Erdmann et al., 2015) or silicate–liquid

55    immiscibility (Dixon & Rutherford, 1979).

56

57    Significantly, plagiogranites have compositional similarities to trondhjemite, tonalite and

58    granodiorite (TTG) rocks that are common in Archean terranes from $4.0 – 2.5$ Ga (e.g.,

59    Drummond et al., 1996; Kerrich & Polat, 2006; Moyen & Martin, 2012; Kusky et al., 2013).

60    Although themselves controversial, Archean TTG are considered, by many, to be generated

61    by the partial melting of mafic igneous source regions (e.g., Drummond et al., 1996; Foley et

62    al., 2002; Rapp et al., 2003; Martin et al., 2005; Moyen & Stevens, 2006; Nutman et al.,

63    2009; Hastie et al., 2015, 2016). Significantly, the compositional similarity of Phanerozoic

64    oceanic plagiogranites to Archean TTG suggests that if we can better understand how

65    plagiogranites are formed, it may further our understanding of how primitive continents were

66    formed on the early Earth (Rollinson, 2008, 2009, 2014).

67

68    In this paper, we present major and trace element data for oceanic plagiogranites sampled

69    from a sheeted dyke complex within the Late Cretaceous (Neo-Tethyan) Muslim Bagh

70    Ophiolite in north-western Pakistan (Kakar et al., 2012). We investigate the composition of

71    these plagiogranitic lenses and dykes in the sheeted dyke complex to determine their

72    petrogenesis. We then discuss the implications of these results for the generation of Archean

73    continental crust.

74

75    **2. Ophiolites and plagiogranites**

76    Oceanic plagiogranites are found throughout geological time, in both the Precambrian (e.g.,

77    Samson et al., 2004; Kaur & Mehta, 2005) and Phanerozoic (e.g., Tilton et al., 1981; Flagler

78    & Spray, 1991; Rollinson, 2009), and are common in the crustal sections of ophiolitic

79    sequences (e.g., Flagler & Spray, 1991; Amri et al., 1996; Twining, 1996; Yaliniz et al.,

80    2000; Samson et al., 2004). Plagiogranites have also been recovered from recent oceanic

81    ridge systems around the world, for example, the Southwest Indian (e.g., Dick et al., 2000),

82    Central Indian (e.g., Nakamura et al., 2007) and Mid-Atlantic Ridges (e.g., Aranovich et al.,

83    2010; Grimes et al., 2011). The morphology of oceanic plagiogranites is complex and they

84    have been documented in a range of intrusive forms, from small veins (millimetre- to

85    centimetre-scale; e.g., Dick et al., 2000; Nakamura et al., 2007), to dykes and inclusions

86    (millimetre- to metre-scale; e.g., Flagler and Spray, 1991; Jafri et al., 1995), to large

87    kilometre-scale plutonic bodies (e.g., Rollinson, 2009).

88

89    Oceanic plagiogranites are predominantly composed of sodic plagioclase and quartz, with

90    mafic (usually hornblende and pyroxene) minerals being minor constituents (<10%), and K-

91    feldspar being a rare phase. In addition to the major modal mineralogy, several accessory

92    minerals including zircon, magnetite and ilmenite are also commonly found in oceanic

93    plagiogranites (Coleman & Peterman, 1975; Coleman & Donato, 1979).

94

95    In the mid-1970s, plagiogranites were considered to represent the likely silicic end products

96    of crystallising basaltic magmas (Coleman & Peterman, 1975; Coleman & Donato, 1979).

97   Although such a crystallisation model is still advocated by some authors, who have shown

98   that oceanic plagiogranites fall along the liquid lines of descent of evolving magmas in other

99   ophiolite units (e.g., Jafri et al., 1995; Rao et al., 2004; Freund et al., 2014), the genesis of

100  oceanic plagiogranites is more commonly attributed to the partial melting of mafic igneous

101  source regions (Gerlach et al., 1981; Flagler & Spray, 1991; see Koepke et al., 2007 for a

102  review of oceanic plagiogranite petrogenesis models). Melting models propose that oceanic

103  plagiogranites are derived through partial melting of mafic protoliths; either by hydrous

104  partial melting of crustal gabbros (e.g., Gerlach et al., 1981; Flagler & Spray, 1991; Amri et

105  al., 1996) or the assimilation and partial melting of hydrothermally altered sheeted dykes

106  (e.g., Gillis & Coogan, 2002; France et al., 2009, 2010; Erdmann et al., 2015).

107

108  A partial melting origin is supported by the experimental work of Koepke et al. (2004), who

109  undertook hydrous melting experiments on oceanic cumulate gabbros at temperatures from

110  900 – 1060 °C and a relatively shallow pressure of 0.2 GPa. Koepke et al. (2004) showed that

111  lower temperature runs (900 – 940 ℃) generated partial melts with similar major element

112  compositions to natural oceanic plagiogranites. One important finding from the P-T

113  experiments was that the melts replicate the low $TiO_2$ concentrations that can be found in

114  oceanic plagiogranites (<1 wt.%; Koepke et al., 2004). Low $TiO_2$ is now considered a key

115  characteristic of oceanic plagiogranites that have been derived by partial melting, as opposed

116  to oceanic plagiogranites derived through fractional crystallisation that display higher $TiO_2$

117  contents (>1 wt.%; Koepke et al., 2004, 2007). Further experimental work conducted by

118  France et al. (2010) has also shown that oceanic plagiogranites derived by partial melting

119  have low TiO2 contents, supporting the experimental work of Koepke et al. (2004).

120

121  **3. Geological setting**

## 3.a. Regional setting

123    The Muslim Bagh Ophiolite (MBO) is one of a number of ophiolites (i.e., Bela, Waziristan,

124    Khost, Zhob) of Neo-Tethyan origin (Kakar et al., 2014) that comprise the Western Ophiolite

125    Belt of the Zhob Valley, north-western Pakistan (Ahmad & Abbas, 1979; Mahmood et al.,

126    1995; Gnos et al., 1997) (Fig. 1). These ophiolites represent fragments of Neo-Tethyan Ocean

127    crust that were obducted onto the margin of the Indian continent prior to its final collision

128    with Asia (e.g., Gnos et al., 1997; Khan et al., 2009) and, therefore, they mark the boundary

129    between the Indian and Eurasian Plates (Asrarullah et al., 1979; Mengal et al., 1994; Gnos et

130    al., 1997).

131

132    The Muslim Bagh area comprises four main geological units (Fig. 1). These units are (south

133    to north) the Indian Passive Margin, the Bagh Complex, the MBO and the Flysch Belt

134    (Mengal et al., 1994; Kakar et al., 2014). Triassic to Palaeocene sediments of the Indian

135    Passive Margin (Kakar et al., 2014) are overthrust by the Mesozoic Bagh Complex along the

136    Gawal Bagh thrust (Mengal et al., 1994). The Bagh Complex comprises a series of thrust

137    bounded units, including a melange unit, two volcanic units (basalt-chert unit [Bbc],

138    hyaloclastite-mudstone unit [Bhm]) and a sedimentary unit (Bs) [see Mengal et al. (1994) for

139    detailed descriptions of each unit]. Thrusted over the Bagh Complex is the MBO (Kakar et

140    al., 2014), described in more detail below. The uppermost unit is the Eocene to Holocene

141    Flysch Belt that rests unconformably on top of the MBO and Bagh Complex in the Katawaz

142    Basin (Mengal et al., 1994; Qayyum et al., 1996; Kasi et al., 2012). The Flysch Belt can be

143    broadly divided into four thrust bounded formations (Nisai, Khujak, Multana and Bostan

144    formations) comprising fluvial and deltaic successions (Qayyum et al., 1996; Kasi et al.,

145    2012).

146

147 **3.b. Muslim Bagh Ophiolite**

148 The MBO is exposed as two massifs, the Saplai Tor Ghar and Jang Tor Ghar Massifs

149 (Ahmad & Abbas, 1979; Mahmood et al., 1995; Gnos et al., 1997) (Fig. 1). The tectonic

150 setting of formation of the MBO has been variously interpreted as a mid-ocean ridge

151 (Mahmood et al., 1995), a back-arc basin (Siddiqui et al., 1996) or an island arc (M. Khan et

152 al., 2007). However, most recently Kakar et al. (2014) have presented evidence that the MBO

153 formed above a slow spreading supra-subduction zone, based on both the structure of the

154 ophiolite and its arc-like geochemistry. Recent U-Pb dating of zircons in MBO plagiogranites

155 by Kakar et al. (2012) gave a crystallisation age of $80.2 \pm 1.5$ Ma that is similar to ~82-81 Ma

156 K-Ar ages obtained by Sawada et al. (1995). Dating of amphiboles from the sub-ophiolitic

157 metamorphic sole have yielded K-Ar and plateau Ar/Ar ages of $80.5 \pm 5.3$ Ma (Sawada et al.,

158 1995) and $70.7 \pm 5$ Ma (Mahmood et al., 1995), respectively. The younger age of $70.7 \pm 5$

159 Ma (Mahmood et al., 1995) is interpreted to date the age of emplacement of the MBO which,

160 when taken in conjunction with the crystallisation age of the ophiolite, suggests that the

161 ophiolite was obducted soon after formation (e.g., Kakar et al., 2014).

162

163 The Saplai Tor Ghar Massif displays a near-complete ophiolite sequence (Kakar et al., 2014),

164 with only the extrusive basalts absent (Mahmood et al., 1995). The Jang Tor Ghar Massif

165 however, only preserves mantle sequence rocks (i.e., foliated peridotite) of the oceanic

166 lithosphere (Mahmood et al., 1995; Kakar et al., 2014). The mantle sequence of the MBO has

167 been divided into a foliated peridotite section and mantle-crust transition zone (Kakar et al.,

168 2014). The foliated peridotite is located in both massifs, and comprises serpentinised

169 harzburgite with minor dunite and chromite deposits (Mahmood et al., 1995; M. Khan et al.,

170 2007; Kakar et al., 2014). Lherzolite is also found in the lower part of the mantle sequence

171 (Kakar et al., 2014). The mantle-crust transition zone of the MBO is a dunite-rich zone with

172    minor gabbro, wherlite, pyroxenite and chromite only exposed in the Saplai Tor Ghar Massif

173    (Mahmood et al., 1995; M. Khan et al., 2007; S. D. Khan et al., 2007; Kakar et al., 2014).

174    Chromite bodies of the transition zone are larger than those in the foliated peridotite section

175    of the mantle sequence (Kakar et al., 2014).

176

177    The oceanic crustal sequence, as exposed in the Saplai Tor Ghar Massif, comprises a 200 –

178    1500 m thick ultramafic-mafic cumulate zone (Ahmad & Abbas, 1979; Siddiqui et al., 1996)

179    and a 1 km thick, poorly developed sheeted dyke complex (Siddiqui et al., 1996; M. Khan et

180    al., 2007). The ultramafic-mafic cumulate zone displays both single and cyclic sequences

181    grading from basal dunite, through pyroxenite, to gabbro, with infrequent anorthosite at the

182    top of the cumulate zone (Ahmad & Abbas, 1979; Siddiqui et al., 1996; Kakar et al., 2014;

183    M. Khan et al., 2007; Kakar et al., 2014). Above the cumulate zone, the sheeted dykes are

184    doleritic, dioritic and plagiogranitic in composition and all display greenschist to amphibolite

185    grade metamorphism (Sawada et al., 1995; Kakar et al., 2014).

186

187    Plagiogranites of the MBO are exclusively located at the base and middle portions of the

188    sheeted dyke complex (Mahmood et al., 1995; Siddiqui et al., 1996). The plagiogranites are

189    rare, comprising <5% by volume of the sheeted dyke complex, and take the form of dykes

190    and small lenses (Fig. 2). They are discontinuous, intrusive bodies, sometimes tapering,

191    displaying a range of sizes. Lenses range from 0.1×0.3 m to 1.0×3.0 m, with more dyke-like

192    bodies ranging from 0.3×1.0 m to 1.5×3.0 m. The plagiogranites have sharp contacts with the

193    enclosing sheeted dykes, and have also undergone greenschist-amphibolite facies

194    metamorphism with foliated to mylonitised textures (Sawada et al., 1995; Siddiqui et al.,

195    1996; Kakar et al., 2014). Samples for the current study were collected from a range of

196    separate plagiogranite dykes and lenses from across the region. The general sampling locality

197    is shown on Figure 1 with more detailed localities and sample information given in online

198    Supplementary Material A at http://journals.cambridge.org/geo.

199

200    **4. Petrography**

201    The plagiogranites sampled from the MBO for the current study are predominately composed

202    of quartz (~40 vol.%) and plagioclase (~50 vol.%), with hornblende and pyroxene

203    comprising minor amounts (<<10 vol.%; hornblende > pyroxene), and zircon and Fe-Ti

204    oxides common as accessory phases. Phenocryst phases of plagioclase, quartz, hornblende

205    and pyroxene are surrounded by a fine groundmass composed of plagioclase, quartz,

206    hornblende, pyroxene, potassium feldspar (rare), and accessory phases. All phenocryst phases

207    have sub-hedral to anhedral crystal shapes, with plagioclase displaying simple and albite

208    twinning, while hornblende twinning is rare. Throughout the sections, quartz is composed of

209    sub-grains. However, unlike Coleman & Peterman's (1975) original definition of oceanic

210    plagiogranites, the MBO plagiogranites do not display vermicular intergrowths of quartz and

211    plagioclase. Evidence for hydrothermal alteration and low-grade metamorphism includes

212    moderate sericitisation of plagioclase crystals (concentrated in the core of crystals; Fig. S1,

213    online Supplementary Material B at http://journals.cambridge.org/geo).

214

215    **5. Geochemical results**

216    **5.a. Analytical techniques**

217    Plagiogranite samples were prepared and analysed for major, minor and trace elements at the

218    School of Earth and Ocean Sciences, Cardiff University, Wales, U.K. Loss on ignition (LOI)

219    was measured using ~1.5 ± 0.0001g of sample powder baked at 900°C in a Vecstar Furnace

220    for 2 hours. Major and minor elements and Sc were measured using a JY-Horiba Ultima 2

221    Inductively Coupled Plasma Optical Emission Spectrometry (ICP-OES). Minor, trace and the

222   rare earth elements (REE) were measured using a thermoelemental X series (X7) Inductively

223   Coupled Plasma Mass Spectrometry (ICP-MS) following methods described by McDonald

224   and Viljoen (2006). Accuracy and precision of the data were assessed using the international

225   standard reference materials JB1a, JA2 and JG-3 (obtained analysis, certified values and

226   detection limits for JB1a are shown in Table S1, online Supplementary Material C at

227   http://journals.cambridge.org/geo). The full data set of plagiogranite samples are shown in

228   Table 1.

229

230   **5.b. Element mobility**

231   The altered nature of the plagiogranite samples means that some of the major elements and

232   large ion lithophile elements (LILE) may have been mobilised relative to the high field

233   strength elements (HFSE) and REE (e.g., Hastie et al., 2007). Although low LOI values (0.59

234   – 2.45 wt. %) suggest that the plagiogranites have suffered little alteration, the high

235   proportion of quartz (~40%) means that the effective LOI of the non-quartz components may

236   double the whole rock values. However, major element (vs. LOI) variation plots of the

237   plagiogranite samples show no correlation with LOI, all displaying very low $R^2$ values (see

238   Fig. S1 of Supplementary Material C at http://journals.cambridge.org/geo). With the

239   exception of MgO (<0.52), all major elements display $R^2$ values of <0.32. These data suggest

240   that the major element concentrations are not primarily controlled by alteration, and can

241   confidently be used to compare to literature Archean TTG data. Additionally, Sr (vs. LOI;

242   Fig. S2, online Supplementary Material C at http://journals.cambridge.org/geo) also displays

243   a very low $R^2$ value of <0.45. Consequently, the following discussion concentrates on the

244   major elements and HFSE and REE, generally regarded as relatively immobile up to

245   greenschist facies (e.g., Floyd & Winchester, 1975; Pearce & Peate, 1995; Hastie et al., 2007,

246   2009).

247

248 **5.c. Major elements**

249 The plagiogranites display a relatively narrow, high-$SiO_2$ range [70.8 – 80.2 wt.% (anhydrous

250 values)], with most also having relatively high $Al_2O_3$ (10.7 – 15.8 wt.%) and $Na_2O$ (1.7 – 4.5

251 wt.%) contents (Fig. 3). Samples have low $TiO_2$ (<0.4 wt.%), MgO (0.1 – 1.8 wt.%) and $K_2O$

252 (<1.1 wt.%). $Al_2O_3$, MnO (not shown), MgO and $K_2O$ decrease with increasing $SiO_2$, while

253 other oxides, such as $TiO_2$, $Na_2O$, $Fe_2O_{3(T)}$ and CaO show little to no correlation (Fig. 3).

254 Also, the plagiogranites do not fall on clear liquid lines of descent along with the gabbros and

255 sheeted dykes of the MBO. On a normative ternary An-Ab-Or plot, the plagiogranites

256 classify as tonalites and trondhjemites (Fig. 4).

257

258 The major element abundances of the plagiogranites are very similar to those of Archean

259 TTG (Condie, 2005; Martin et al., 2005; Moyen and Martin, 2012); with TTG compositions

260 consistently plotting at the lower $SiO_2$ end of the plagiogranite compositions (Fig. 3).

261 However, this similarity is not observed in $K_2O$ contents, with TTG generally having much

262 higher $K_2O$ contents (1.65 – 2.22 wt. %) compared to the MBO plagiogranites (<1.1 wt. %).

263

264 **5.d. Trace elements**

265 The plagiogranites show no convincing intra-formation fractionation trends on trace element

266 variation plots (Fig. 5). This is not surprising considering that the samples are collected from

267 a diverse range of geographically distinct dykes and lenses. The plagiogranites span a wide

268 range in Zr concentrations (~20 – 280 ppm); however, the majority of samples fall in the

269 range 20 – 90 ppm, with only three having higher concentrations (130, 199, 283 ppm)

270 suggestive of zircon accumulation (e.g., Rollinson, 2009). In general, the plagiogranites have

271 lower trace element concentrations than the sheeted dyke complex of the MBO and, with the

11

272   exception of Sr, have trace element contents similar to, or slightly greater than, the majority

273   of the gabbros of the crustal section of the ophiolite (Fig. 5). As is the case with the major

274   elements (Fig. 3), the plagiogranites also do not fall on clear liquid lines of descent along

275   with the gabbros and sheeted dykes of the MBO (Fig. 5). As seen above, the major element

276   compositions of the MBO plagiogranites are very similar to those of TTG compositions (Fig.

277   3); however, this similarity is not as evident in the trace elements (Fig. 5).

278

279   The plagiogranites show broadly coherent trends in the middle- to heavy-REE (M/HREE) on

280   chondrite-normalised REE plots, but have variable light-REE (LREE) contents, from

281   markedly enriched to relatively depleted patterns [e.g., $4.8 - 0.7$ $(La/Sm)_N$] (Fig. 6a, c). The

282   LREE enriched patterns shown by the majority of the plagiogranite samples are inconsistent

283   with the original definition of plagiogranites (Coleman & Peterman, 1975), and are shown to

284   be enriched relative to the well-studied crustal plagiogranites from the Oman and Troodos

285   Ophiolites (Fig. 6a). However, plagiogranites from the Sjenica (Milovanovic et al., 2012) and

286   Tasriwine Ophiolites (Samson et al., 2004) have recently been reported that have LREE

287   enriched patterns (Fig. 6a). When compared to Archean TTG compositions, plagiogranite

288   samples are mostly shown to not be as enriched in the LREE (Fig. 6a). Most samples also

289   show a slight chondrite normalised enrichment in the heaviest REE relative to the MREE and

290   display small U-shaped (concave upwards) patterns. The U-shaped patterns can be quantified

291   using the Dy/Dy* ratio of Davidson et al. (2012), which ranges from $0.96 - 0.43$ (Fig. 6b).

292   Most plagiogranites have weak positive Eu anomalies [$1.06 - 1.51$ (Eu/Eu)*], with only three

293   samples having negative Eu anomalies [$0.74 - 0.94$] (Fig. 6c). Interestingly, two of the three

294   samples with negative Eu anomalies are also significantly enriched in the LREE.

295

296  On normal mid-ocean ridge basalt (N-MORB) normalised multi-element plots, most

297  plagiogranites display relatively flat patterns at concentrations just below N-MORB, with

298  positive Th anomalies and negative Nb-Ta-Ti anomalies (Fig. 7a). Zr and Hf contents vary

299  from enriched to depleted, relative to N-MORB. Most samples also have positive Sr

300  anomalies; however three samples have negative Sr anomalies, two of which display

301  corresponding negative Eu anomalies (Fig. 6c).

302

303  **6. Discussion**

304  The modal abundance of quartz and plagioclase in combination with the low $K_2O$ contents

305  (<1.1 wt.%) of the MBO plagiogranites is similar to oceanic plagiogranites found elsewhere

306  (e.g., Gerlach et al., 1981; Amri et al., 1996; Rollinson, 2009). Additionally, the trace

307  element compositions of plagiogranites from the Muslim Bagh, Oman and Troodos

308  Ophiolites all show a high degree of compositional overlap (Fig. 7a, plagiogranite field)

309  (Rollinson et al., 2009; Freund et al., 2014). Nevertheless, the LREE-enriched and slightly

310  concave-upward MREE patterns of the majority of MBO samples are distinct relative to the

311  original oceanic plagiogranite definition (Coleman & Peterman, 1975; Coleman & Donato,

312  1979).

313

314  High $SiO_2$ (>70 wt.%) and $Na_2O$ (3< $Na_2O$ <4.5 wt.%) concentrations and low modal K-

315  feldspar contents, low $K_2O/Na_2O$ ratios and low $Fe_2O_3+MgO+MnO+TiO_2$ (most <5 wt.%) of

316  the MBO plagiogranites make them compositionally similar to Archean TTG as defined by

317  Martin et al. (2005) and Moyen and Martin (2012). Additionally, when compared to Archean

318  TTG compositions on an N-MORB normalised multi-element plot, the plagiogranites display

319  broadly similar concentrations, overlapping the TTG field at the lower LREE and higher

320  HREE concentrations (Fig. 7b).

321

**6.a. Plagiogranite petrogenesis**

323     The majority of plagiogranites display enrichment in the LREE relative to the HREE (Fig.

324     6c) and all plagiogranites have negative Nb-Ta and positive Th anomalies (Fig. 7a).

325     Additionally, the N-MORB-like concentrations of the other trace elements suggest that the

326     plagiogranites (Fig. 6c) were generated at a MOR setting with a subduction input, likely a

327     supra-subduction zone. This supports recent work by Kakar et al. (2014) who propose a

328     supra-subduction model for the formation of the MBO. However, the petrogenesis of oceanic

329     plagiogranites is controversial with fractional crystallisation, partial melting or silicate –

330     liquid immiscibility being variously proposed as petrogenetic models [see Koepke et al.

331     (2007) for a review]. Below, we discuss the implications the plagiogranite compositions have

332     for each of the possible petrogenetic models.

333

334     *6.a.1. Fractional crystallisation and liquid immiscibility*

335     The layered gabbros and sheeted dykes of the MBO crustal section represent possible

336     cumulates and parental melts, respectively from which to derive the plagiogranites by

337     crystallisation. However, major and trace element variation diagrams (Fig. 3**,** 5) show that the

338     plagiogranites do not plot along the same liquid lines of descent as any of the other MBO

339     units. The fact that the plagiogranites define their own distinct field clearly indicates that they

340     are not related to the other units by simple fractional crystallisation processes. The lack of

341     intermediate units within the ophiolite sequence also argues against an origin for the

342     plagiogranites by fractional crystallisation from a basic parental melt. Additionally, the

343     narrow $SiO_2$ range of the plagiogranites would suggest fractional crystallisation did not play a

344     primary role in their petrogenesis.

345

346    Concave-upwards patterns displayed by the plagiogranites (Fig. 6c) support a role for

347    amphibole during their petrogenesis; a result of amphiboles preference for the MREE over

348    the LREE and HREE (e.g., Davidson et al., 2012). However, the concave-upward pattern on

349    its own does not indicate whether amphibole was crystallising from a parental magma or

350    acting as a residual phase during the fusion of a mafic protolith.

351

352    An origin by silicate-liquid immiscibility (e.g., Dixon & Rutherford, 1979) is also unlikely

353    for the MBO plagiogranites. This is evidenced by the absence of the associated immiscible

354    Fe-rich liquid (as Fe-rich mafic units) from the MBO.

355

356    *6.a.2. Partial melting*

357    Experimental work of Koepke et al. (2004) and France et al. (2010) has shown that low $TiO_2$

358    contents (<1 wt.%; Koepke et al., 2004) are characteristic of oceanic plagiogranites derived

359    through partial melting a mafic protolith; a consequence of the gabbroic protoliths having

360    initially low $TiO_2$ contents, typical of cumulate gabbros of the oceanic crust (Koepke et al.,

361    2004, 2007). Low $TiO_2$ contents of the MBO plagiogranites (Fig. 3b) are similar to those in

362    the experimentally derived high-$SiO_2$ melts of Koepke et al. (2004), suggesting they were

363    derived by partial melting of a gabbroic protolith in the crustal sequence of the MBO. In

364    addition, $TiO_2$ contents of the MBO plagiogranites plot below the boundary line drawn by

365    Koepke et al. (2007) that separates plagiogranites derived by hydrous partial melting (plot

366    below black dashed line, Fig. 3b) from those plagiogranites derived by crystallisation or

367    immiscibility processes (plot above black dashed line).

368

369    Additionally, as shown in Figure 3, major element concentrations of the MBO plagiogranites

370    are similar to Archean TTG (Condie, 2005; Martin et al., 2005; Moyen and Martin, 2012),

371    which are generally regarded to have been generated through partial melting of a mafic

372    igneous protolith (e.g., Drummond et al., 1996; Foley et al., 2002; Rapp et al., 2003; Martin

373    et al., 2005; Moyen & Stevens, 2006; Nutman et al., 2009; Hastie et al., 2015, 2016). We

374    suggest that the lower $K_2O$ contents displayed by the plagiogranites, compared to Archean

375    TTG, is the result of the TTG rocks being derived from a more primitive mantle prior to

376    continental crust extraction, and therefore a less depleted mantle than the present. Trace

377    element variation plots (Fig. 5) however do not show as convincing a similarity between the

378    MBO plagiogranites and Archean TTG as do the major element variation plots (Fig. 3).

379    Nevertheless, overall the MBO plagiogranites have broadly similar trace element

380    compositions to Archean TTG (Fig. 7b).

381

382    Negative Eu and Sr anomalies (Fig. 6c, 7a) and decreasing $Al_2O_3$ with increasing $SiO_2$ (Fig.

383    3) in some samples could potentially be explained by a small amount of late stage plagioclase

384    fractional crystallisation. However, negative Eu and Sr anomalies can also be the result of

385    plagioclase in the melting residue, while the decrease in $Al_2O_3$ with $SiO_2$ can be reproduced

386    through small degrees of partial melting as demonstrated by Beard & Lofgren (1991). In the

387    following section we will use trace element modelling to test a partial melting model for the

388    MBO plagiogranites.

389

390    **6.b. Modelling of partial melting**

391    To model the partial melting of a mafic protolith the non-modal batch melting equation of

392    Shaw (1970) was used for the calculations:

393

$$C_l = \frac{C_0}{D_0 + F(1 - P)} \qquad [1]$$

394

395    where, $C_1$ is the concentration of a particular trace element in a resultant melt, $C_0$ is the

396    concentration of an element in the source region prior to partial melting, F is the mass

397    fraction of melt generated, $D_0$ is the bulk partition coefficient of an element prior to partial

398    melting and P is the partition coefficient of an element weighted by the proportion

399    contributed by each mineral phase to the melt. Hornblende gabbro, C51 [from Kakar et al.

400    (2014)] was used as the protolith. This sample was collected from the cumulate sequence of

401    the crustal section of the MBO and was chosen as the protolith since the concave-upward

402    pattern shown by the plagiogranites suggests that amphibole may have been left behind in the

403    melting residue. The partition coefficients used are those for elements in equilibrium with

404    TTG-like silicic melts from Bedard (2006). Mineral modes of the hornblende gabbro are

405    those of Kakar et al. (2014) and Siddiqui et al. (1996). Melt modes were calculated using 1

406    kbar experimental runs from Beard & Lofgren (1991) as they provide enough petrological

407    information to carry out the calculation. Melting was stopped at 14.5%, as this is the point at

408    which hornblende is exhausted from the protolith. Mineral and melt modes, partition

409    coefficients, hornblende gabbro starting composition and model results can be found in Table

410    S1, online Supplementary Material D at http://journals.cambridge.org/geo.

411

412    Figure 7c shows that the incompatible trace element patterns (including negative Nb and Ti

413    anomalies and positive Th and Zr anomalies) of the plagiogranites can be replicated by 5 –

414    10% partial melting of the hornblende gabbro. Nonetheless the modelling generates a larger

415    negative Sr anomaly than seen in the MBO plagiogranites. This result is attributed to the use

416    of a high Sr partition coefficient in plagioclase (6.65; Bedard, 2006) and this discrepancy can

417    be removed if a lower partition coefficient is used [i.e., 3, based on the range reported by

418    Laurent et al. (2013)].

419

420  Despite the evidence supporting a partial melting model for the MBO plagiogranites, the

421  reason behind the negative $K_2O$ trend displayed by the plagiogranites when plotted against

422  $SiO_2$ (Fig. 3) is uncertain. It is possible however that the negative trends displayed by both

423  $K_2O$ and $Al_2O_3$ are the result of an interplay between fractional crystallisation (plagioclase

424  and biotite(?)) and/or varying degrees of partial melting and source variation.

425

**6.c. Comparison with other Tethyan Ophiolite plagiogranites and implications for the tectonomagmatic setting of the Muslim Bagh Ophiolite**

428  As we have shown, some geochemical characteristics of the MBO plagiogranites (i.e., LREE

429  enriched and concave-upward MREE patterns) do not conform to the definition of oceanic

430  plagiogranites as proposed by Coleman & Peterman (1975). The results from this study are

431  similar to previous plagiogranite analyses from the MBO presented by Kakar et al. (2014),

432  who also report MBO plagiogranites with LREE enriched patterns $[1 - 7, (La/Sm)_N]$, as well

433  as negative Nb-Ta-Ti anomalies and low $TiO_2$ contents ($\leq 0.20$ wt. %).

434

435  The MBO plagiogranites are significantly different to those from other Tethyan Ophiolites in

436  terms of both field and geochemical characteristics. First, LREE contents of Troodos and

437  Oman Ophiolite crustal plagiogranites are relatively depleted compared to the HREE (Fig.

438  6a) (Rollinson et al., 2009; Freund et al., 2014) and therefore a more depleted source is

439  required for these plagiogranites relative to the MBO plagiogranites. It is however beyond the

440  scope of this study to investigate further the difference in source enrichment between the

441  MBO plagiogranites and those plagiogranites situated in the Oman and Troodos Ophiolites.

442  Secondly, the plagiogranites of the MBO are solely located in the crustal section of the

443  ophiolite, whereas geochemically distinct groups of plagiogranites have been identified in

444  crust and mantle sections of the Troodos and Oman Ophiolites (Rollinson, 2009, 2014;

445  Freund et al., 2014). Thirdly, the MBO plagiogranites are generally smaller intrusive bodies

446  (on a scale of no more than a few meters) than those found in both the Troodos and Oman

447  Ophiolites, where plagiogranites range from several tens of meters to kilometre sized plutons

448  (Rollinson et al., 2009; Freund et al., 2014).

449

450  The poorly developed sheeted dyke complex (M. Khan et al., 2007; Kakar et al., 2014) of the

451  MBO crustal section is likely the result of the imbalance between spreading rate and magma

452  supply in a supra-subduction zone tectonic setting (Robinson et al., 2008). Robinson et al.

453  (2008) have proposed that both the forearc and backarc of a supra-subduction zone generally

454  experience lower magma supply rates, due to eruptions at the volcanic arc, and high

455  extensional strain rates. Therefore, the small size, restricted distribution and lack of

456  geochemical variability (i.e., uniform composition) amongst the MBO plagiogranites could

457  be a result of this decreased magma supply in the supra-subduction zone where the MBO

458  crystallised. Consequently, the decreased magma supply results in a small degree of partial

459  melting of the plagiogranite source (i.e., crustal hornblende gabbros).

460

461  **6.d. Implications for Archean TTG genesis**

462  Most previous and current research into Archean TTG petrogenesis favours models in which

463  juvenile Archean continental crust is generated by partial melting of mafic igneous protoliths

464  (e.g., Sen & Dunn, 1994; Wolf & Wyllie, 1994; Foley et al., 2002; Rapp et al., 2003; Moyen

465  & Stevens, 2006; Laurie & Stevens, 2012; Zhang et al., 2013; Ziaja et al., 2014; Hastie et al.,

466  2016), the setting of which is still controversial, with both subduction/flat slab

467  subduction/underthrusting (e.g., Drummond et al., 1996; Martin et al., 2005; Nutman et al.,

468  2009; Hastie et al., 2015) and intracrustal (Hamilton, 1998; Hawkesworth et al., 2016)

469  settings having been proposed for the derivation of Archean TTG of various ages.

470

471  Since the original definition of oceanic plagiogranites in the mid-1970s by Coleman &

472  Peterman (1975), oceanic plagiogranites have been shown to differ compositionally to

473  Archean TTG; being less potassic, and having MORB-like LREE and flat HREE patterns.

474  Numerous studies on oceanic plagiogranites from the Oman Ophiolite (Rollinson, 2008,

475  2009, 2014) have suggested that although the Oman Ophiolite plagiogranites have

476  compositions that are similar to oceanic plagiogranites [as defined by Coleman & Peterman

477  (1975)] and differ compositionally from Archean TTG, they can be instructive on Archean

478  TTG genesis. Rollinson (2009) noted that in addition to the conditions of plagiogranite

479  petrogenesis, a source region enriched in the LREE is also required in order to generate the

480  LREE-enriched nature of Archean TTG. Additionally, Rollinson (2008) has suggested that

481  trondhjemite (plagiogranite) petrogenesis in the Oman Ophiolite acts as a possible analogue

482  for the generation of Earth's first felsic crust in the Hadean. Rollinson (2008) has argued that

483  early (Hadean) felsic crust was of low volume and this corresponds to the low volume of

484  plagiogranites we see in recent ophiolite sequences.

485

486  The MBO plagiogranites are compositionally different (LREE-enriched and concave-upward

487  MREE patterns) to the original oceanic plagiogranite definition, but are geochemically

488  similar to Archean TTG (e.g., Condie, 2005; Martin et al., 2005; Moyen and Martin, 2012)

489  (Fig. 3, 7b). Consequently, the MBO plagiogranites can be used as a recent (Late Cretaceous)

490  analogue to investigate the formation of some Archean TTG rocks.

491

492  The MBO plagiogranites are found within mafic crust that was formed at a convergent

493  margin – specifically the upper plate above the subduction zone (e.g., Siddiqui et al., 1996,

494  2011; Kakar et al., 2014). The similarity in composition between the MBO plagiogranites and

495  Archean TTG suggests that some of the earliest silicic continental crust may have been

496    derived from melting the overriding plates in primitive subduction-like zones. We

497    acknowledge that there is a contrast in volume between the MBO plagiogranites and Archean

498    TTG; however, we infer that the genesis of these plagiogranites can be instructive on the

499    generation of some, but not all, Archean TTG. In addition, the overall greater enrichment in

500    the LREE relative to the HREE of Archean TTG compared to the MBO plagiogranites

501    suggests that to source a larger portion of Archean TTG requires a slightly more enriched

502    source than that of the MBO plagiogranites (e.g., Rollinson, 2009). Again, this could possibly

503    be due to the extraction of continental crust, and depletion of the mantle over time.

504

505    **7. Conclusions**

506     1. Oceanic plagiogranites of the MBO are exclusively located at the base and middle

507         portions of the sheeted dyke complex, where they form small, intrusive dyke-like

508         bodies and lenses.

509     2. Low $TiO_2$ contents (<0.4 wt. %) in the plagiogranites and a lack of intermediate rocks

510         in the sheeted dyke complex suggest an origin by partial melting of mafic rocks. This is

511         confirmed by batch melt trace element modelling of a crustal hornblende gabbro from

512         the crustal sequence of the MBO. This modelling shows that the plagiogranites can be

513         replicated by 5 – 10% partial melting, possibly with a small degree of late stage

514         fractional crystallisation of plagioclase(?) to account for negative Sr and Eu anomalies

515         and a decrease in $Al_2O_3$ with $SiO_2$.

516     3. The similarity in composition of the MBO plagiogranites with Archean TTG rocks

517         supports the model that some Archean TTG could be generated by partial melting of a

518         mafic protolith, possibly in the overriding plate of a subduction-like zone.

519

520    *Acknowledgements*:

531

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747     **Figure Captions**

748     **Figure 1**. Geological map of the Muslim Bagh area. Inset highlights the location of the

749     Muslim Bagh Ophiolite in north-western Pakistan (modified from Kakar et al., 2014).

750

751     **Figure 2**. Field photographs of the Muslim Bagh Ophiolite plagiogranites. Plagiogranites are

752     exclusively located within the sheeted dyke complex of the ophiolite crustal sequence, where

753     they take the form of dyke-like bodies (**a**), and lenses (**b**).

754

755     **Figure 3**. Major element variation plots (vs. $SiO_2$) of the Muslim Bagh Ophiolite

756     plagiogranites. Also plotted are the sheeted dykes and gabbros of the crustal section of the

757     Muslim Bagh Ophiolite (data from Kakar et al., 2014) and Archean TTG average

758     compositions; C2005 (Condie, 2005), M2005 (Martin et al., 2005) and MM2012 (Moyen &

759     Martin, 2012). The black dashed line in (b) separates plagiogranites derived by hydrous

760     partial melting (below the line) and those plagiogranites derived through differentiation or

761     liquid immiscibility (above the line) (after Koepke et al., 2007).

762

763     **Figure 4.** Normative An-Ab-Or ternary plot. Muslim Bagh Ophiolite plagiogranites classify

764     as either tonalites or trondhjemites. Fields from Barker (1979).

765

766     **Figure 5**. Representative trace element variation plots of the Muslim Bagh Ophiolite

767     plagiogranites. The sheeted dykes and gabbros of the crustal section of the Muslim Bagh

768     Ophiolite (data from Kakar et al., 2014), and Archean TTG average compositions are also

769     plotted; symbols and references as in Figure 3.

770

771  **Figure 6**. a) Plot of $(La/Sm)_N$ vs. $(Gd/Yb)_N$ highlighting the LREE enriched nature of the

772  majority of the Muslim Bagh Ophiolite plagiogranites relative to the depleted Oman

773  (Rollinson, 2009) and Troodos Ophiolites (Freund et al., 2014). Also plotted are LREE

774  enriched plagiogranites from the Sjenica (Milovanovic et al., 2012) and Tasriwine Ophiolites

775  (Samson et al., 2004) and Archean TTG average compositions (symbols and references as in

776  Figure 3). b) Plot of Dy/Dy* vs. Dy/Yb showing the majority of the Muslim Bagh Ophiolite

777  plagiogranites to plot in the concave-upward quadrant (black dotted lines) and follow the

778  amphibole vector (arrow) in figure 4 of Davidson et al. (2012). The plot quantifies the degree

779  of concavity, and supports a role for amphibole in the petrogenesis of the plagiogranites,

780  either as a residual or crystallising phase. c) Chondrite normalised rare earth element plot of

781  the Muslim Bagh Ophiolite plagiogranites. Normalising values after Sun & McDonough

782  (1989).

783

784  **Figure 7**. a) Normal mid-ocean ridge basalt normalised multi-element plot of the Muslim

785  Bagh Ophiolite plagiogranites. Dashed plagiogranites field represents analyses of

786  plagiogranites from the Troodos (Freund et al., 2014) and Oman (Rollinson, 2009)

787  Ophiolites. b) Normal mid-ocean ridge basalt normalised multi-element plot comparing the

788  Muslim Bagh Ophiolite plagiogranites with Archean TTG average compositions (Condie,

789  2005; Martin et al., 2005; Moyen & Martin, 2012). c) Trace element modelling of batch

790  melting. The primitive mantle normalised multi-element plot compares the trace element

791  composition resulting from trace element melt modelling of a crustal hornblende gabbro with

792  the composition of the Muslim Bagh Ophiolite plagiogranites. Plagiogranite compositions

793  can be replicated by 5 – 10% partial melting of a hornblende gabbro. Dashed black lines

794  represent melts derived by partial melting when using a lower (i.e., 3; Laurent et al., 2013)

795    partition coefficient for Sr in plagioclase. Normalising values after Sun & McDonough

796    (1989).

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**Table 1**. Major and trace element analyses of the Muslim Bagh Ophiolite plagiogranites

| Sample | PI-01 | PI-02 | PI-03 | PI-06 | PI-07 | PI-13 | PI-15 | PI-17 | PI-19 | PI-21 | PI-22 | PI-23 | PI-25 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ (wt. %) | 73.08 | 70.71 | 70.26 | 70.00 | 71.66 | 72.06 | 72.40 | 74.06 | 74.85 | 74.69 | 72.56 | 79.73 | 76.85 |
| $TiO_2$ | 0.30 | 0.31 | 0.30 | 0.34 | 0.30 | 0.31 | 0.34 | 0.30 | 0.29 | 0.29 | 0.39 | 0.10 | 0.28 |
| $Al_2O_3$ | 12.98 | 14.83 | 14.87 | 15.50 | 15.63 | 15.03 | 13.37 | 10.50 | 13.42 | 13.79 | 14.41 | 12.27 | 11.68 |
| $Fe_2O_3$ (t) | 2.81 | 3.48 | 3.00 | 2.47 | 2.40 | 2.60 | 2.84 | 4.03 | 3.02 | 2.79 | 4.47 | 1.25 | 3.87 |
| MnO | 0.04 | 0.05 | 0.04 | 0.04 | 0.04 | 0.04 | 0.04 | 0.03 | 0.03 | 0.02 | 0.04 | 0.01 | 0.04 |
| MgO | 1.81 | 1.01 | 0.78 | 0.66 | 0.63 | 0.42 | 0.92 | 1.22 | 0.50 | 0.40 | 0.37 | 0.10 | 0.19 |
| CaO | 3.68 | 2.87 | 3.02 | 3.94 | 3.38 | 4.01 | 5.67 | 4.72 | 4.01 | 3.93 | 3.10 | 1.91 | 1.86 |
| $Na_2O$ | 3.12 | 4.42 | 4.35 | 4.19 | 4.34 | 4.24 | 3.13 | 1.70 | 3.40 | 3.65 | 3.98 | 4.20 | 4.14 |
| $K_2O$ | 0.52 | 0.98 | 0.64 | 0.62 | 0.71 | 0.55 | 0.43 | 0.33 | 0.18 | 0.17 | 0.29 | 1.06 | 0.12 |
| $P_2O_5$ | 0.03 | 0.03 | 0.04 | 0.05 | 0.05 | 0.05 | 0.06 | 0.06 | 0.07 | 0.07 | 0.14 | 0.02 | 0.06 |
| LOI | 1.83 | 1.71 | 1.75 | 1.16 | 1.32 | 0.81 | 1.09 | 2.45 | 0.82 | 0.76 | 1.72 | 0.59 | 0.81 |
| Total | 100.19 | 100.38 | 99.05 | 98.97 | 100.46 | 100.12 | 100.29 | 99.40 | 100.05 | 99.99 | 100.92 | 100.53 | 99.22 |
| Sc (ppm) | 6.7 | 7.8 | 9.0 | 10.7 | 11.8 | 12.5 | 13.4 | 13.0 | 13.4 | 9.4 | 7.3 | 2.9 | 10.5 |
| V | 59 | 64 | 55 | 45 | 38 | 55 | 45 | 48 | 62 | 44 | 49 | 10 | 8 |
| Cr | 13 | 26 | 2 | 13 | 8 | 7 | 80 | 174 | 12 | 21 | 116 | 18 | 10 |
| Co | 8.6 | 14.8 | 13.5 | 7.9 | 5.6 | 7.1 | 9.3 | 13.7 | 6.8 | 7.3 | 6.6 | 2.0 | 14.3 |
| Ni | 15.7 | 50.2 | 22.4 | 78.5 | 10.4 | 8.9 | 8.6 | 10.5 | 25.3 | 5.0 | 89.1 | 10.1 | 14.2 |
| Ga | 10.9 | 11.3 | 10.7 | 11.0 | 10.4 | 11.2 | 11.2 | 8.8 | 11.6 | 12.1 | 15.8 | 10.0 | 12.1 |
| Rb | 3.8 | 6.1 | 3.4 | 2.4 | 2.9 | 2.8 | 4.1 | 8.1 | 2.0 | 1.5 | 2.9 | 20.7 | 1.6 |
| Sr | 160 | 163 | 167 | 170 | 144 | 163 | 159 | 196 | 155 | 155 | 214 | 105 | 91 |
| Y | 13.0 | 13.6 | 11.3 | 17.3 | 18.0 | 14.2 | 15.8 | 17.7 | 19.4 | 16.3 | 13.0 | 17.9 | 22.0 |
| Zr | 40.3 | 198.8 | 46.6 | 94.9 | 42.7 | 51.2 | 20.7 | 35.2 | 80.0 | 57.8 | 283.0 | 61.1 | 129.8 |
| Nb | 1.28 | 1.05 | 1.17 | 0.79 | 0.69 | 0.68 | 0.80 | 0.80 | 0.89 | 1.42 | 4.13 | 5.19 | 0.70 |
| Cs | 0.05 | 0.07 | 0.10 | 0.07 | 0.09 | 0.06 | 0.05 | 0.12 | 0.04 | 0.03 | 0.07 | 0.13 | 0.02 |
| Ba | 68 | 69 | 71 | 79 | 74 | 71 | 78 | 67 | 60 | 80 | 198 | 502 | 59 |
| La | 5.14 | 2.87 | 2.97 | 3.70 | 2.12 | 3.28 | 2.14 | 3.16 | 1.50 | 1.69 | 21.83 | 15.45 | 4.77 |
| Ce | 9.58 | 6.28 | 6.08 | 7.89 | 5.06 | 6.47 | 5.00 | 6.50 | 4.26 | 4.56 | 37.17 | 27.28 | 11.27 |
| Pr | 1.15 | 0.84 | 0.84 | 1.09 | 0.77 | 0.84 | 0.75 | 0.94 | 0.70 | 0.71 | 4.23 | 3.03 | 1.70 |
| Nd | 4.68 | 3.71 | 3.91 | 5.22 | 3.99 | 3.89 | 3.72 | 4.52 | 3.51 | 3.43 | 14.97 | 10.13 | 7.69 |
| Sm | 1.37 | 1.28 | 1.21 | 1.59 | 1.44 | 1.35 | 1.50 | 1.50 | 1.43 | 1.33 | 2.96 | 2.15 | 2.31 |
| Eu | 0.53 | 0.53 | 0.63 | 0.68 | 0.60 | 0.58 | 0.63 | 0.65 | 0.52 | 0.54 | 0.82 | 0.54 | 0.88 |
| Gd | 1.54 | 1.38 | 1.33 | 1.96 | 1.96 | 1.69 | 1.78 | 1.91 | 2.00 | 1.73 | 2.86 | 2.27 | 2.76 |
| Tb | 0.28 | 0.27 | 0.24 | 0.38 | 0.35 | 0.30 | 0.33 | 0.35 | 0.42 | 0.35 | 0.40 | 0.37 | 0.51 |
| Dy | 1.88 | 1.88 | 1.71 | 2.53 | 2.67 | 2.01 | 2.35 | 2.45 | 2.90 | 2.34 | 2.12 | 2.30 | 3.09 |
| Ho | 0.43 | 0.45 | 0.38 | 0.56 | 0.60 | 0.46 | 0.51 | 0.53 | 0.60 | 0.49 | 0.40 | 0.50 | 0.64 |
| Er | 1.41 | 1.38 | 1.10 | 1.74 | 1.81 | 1.42 | 1.61 | 1.68 | 1.85 | 1.46 | 1.18 | 1.57 | 1.84 |
| Tm | 0.24 | 0.26 | 0.19 | 0.30 | 0.31 | 0.24 | 0.28 | 0.27 | 0.31 | 0.27 | 0.19 | 0.30 | 0.33 |
| Yb | 1.58 | 1.75 | 1.22 | 2.04 | 2.03 | 1.46 | 1.76 | 1.80 | 2.07 | 1.68 | 1.24 | 2.15 | 2.13 |
| Lu | 0.24 | 0.30 | 0.19 | 0.32 | 0.32 | 0.22 | 0.26 | 0.27 | 0.31 | 0.27 | 0.21 | 0.34 | 0.33 |
| Hf | 1.42 | 5.81 | 1.48 | 2.90 | 1.37 | 1.65 | 0.72 | 1.15 | 2.08 | 1.57 | 6.72 | 1.43 | 3.04 |
| Ta | 0.11 | 0.07 | 0.14 | 0.10 | 0.06 | 0.06 | 0.07 | 0.07 | 0.06 | 0.10 | 0.29 | 0.47 | 0.04 |
| Pb | 1.34 | 1.39 | 2.48 | 1.88 | 1.41 | 1.80 | 1.27 | 1.17 | 1.11 | 0.64 | 1.82 | 3.45 | 1.62 |
| Th | 2.17 | 0.51 | 0.25 | 0.26 | 0.20 | 0.20 | 0.23 | 0.47 | 0.45 | 0.33 | 11.92 | 4.95 | 0.61 |
| U | 0.30 | 0.10 | 0.11 | 0.09 | 0.07 | 0.06 | 0.05 | 0.12 | 0.09 | 0.09 | 0.93 | 0.95 | 0.13 |

$Fe_2O_3$ (t): total iron

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Figure 1

Figure 2

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835   Figure 3

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873    Figure 6

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886    Figure 7