

1 **U-Pb geochronology and petrogenesis of peraluminous granitoids from northern Indian**
2 **plate in NW Pakistan: Andean type orogenic signatures from the early Paleozoic along the**
3 **northern Gondwana**

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18 **Abstract**

19 The pre-Himalayan peraluminous magmatic event along the northern margin of Indian plate in
20 north-western Pakistan has been investigated leading to a correlation with the magmatic evolution
21 in other Himalayan and northern Gondwana regions. The two mica granites from Utla and Mansehra
22 regions of NW Pakistan are dominantly megacrystic, strongly peraluminous ($A/CNK > 1.1$) and
23 intruded by aplitic dykes and quartz rich veins. U-Pb zircon dating by SIMS reveals their
24 emplacement during the early Paleozoic, ranging from 476 Ma to 480 Ma. These granites are
25 enriched in light rare-earth elements (LREEs) and show similar REE patterns with negative Eu
26 anomalies. Geochemical modelling indicates that these granites were derived mainly from the
27 partial melting of pelitic sources followed by the evolution of melt via fractional crystallization of
28 feldspars, biotite, muscovite, apatite, and/or zircon, with the aplite dykes representing the very last
29 fractionation product. Based on their compositions, source rock characteristics and U-Pb
30 geochronology, we assign these to the regional association of other Cambro-Ordovician granitoids
31 from the Himalayas and northern Gondwana terranes. Due to these similarities alongside other
32 metamorphic, stratigraphic and geochemical evidence, an early Paleozoic Andean-type orogenic
33 event is proposed for the genesis of these granitoids where the process could have been initiated by
34 the subduction of the Proto-Tethys oceanic lithosphere beneath the northern Gondwana continental
35 margin.

36 **Key Words:**

37 U-Pb Geochronology; Geochemistry; Early Paleozoic Utla granites; Andean-type orogeny; Northern
38 Indian plate; NW Pakistan

39 **1. Introduction**

40 The northern Indian plate has experienced multiple tectonic regimes during its geological history. In
41 the early Cenozoic, it records convergent tectonic conditions during the Himalayan orogenic
42 episodes initiated by the intra-oceanic subduction of the Tethys Ocean (Searle et al., 2009; Jagoutz

43 et al., 2011; Burg, 2011). Preceding this, it had been subjected to extensional settings during the late
44 Paleozoic as indicated by the presence of rift-related alkaline lithologies (Kempe and Jan, 1970;
45 Garzanti et al., 1999; Noble et al., 2001). Recent research has highlighted the role of magmatism as
46 part of widespread tectonothermal activity during the early Paleozoic of the Himalaya (Gehrels et al.,
47 2006a, 2006b; Cawood et al., 2007) as well as the other regions of northern Gondwana (DeCelles et
48 al., 1998, 2000, 2004; Booth et al., 2004; Hu et al., 2013; Wang et al., 2013; Ding et al., 2015; Li et al.,
49 2016).

50 The mechanism of early Paleozoic events are poorly understood as pre-Himalayan events are
51 commonly overprinted by pervasive and extreme Tertiary tectonism. Pre-Himalayan rocks are
52 generally associated with Pan-African orogenesis in response to the final accretion of the Gondwana
53 continents (Murphy and Nance, 1991; Meert and Van der Voo, 1997; Miller et al., 2001; Liu et al.,
54 2012; Yang et al., 2012). In contrast, an Andean type orogeny has been proposed for these rocks due
55 to the subduction of proto-Tethyan oceanic lithosphere under the northern Gondwana continents
56 (Cawood et al., 2007; Dong et al., 2013; Spencer et al., 2012; Hu et al., 2013; Wang et al., 2013; Li et
57 al., 2016).

58 Research on the northern margin of the Indian plate has commonly focussed on the Cenozoic
59 Himalayan uplift and related tectonism, metamorphism and magmatism, however, the pre-
60 Himalayan details are relatively scarce. In this paper, we aim to help fill this knowledge gap by using
61 a comprehensive geochemical and geochronological dataset to improve the understanding of pre-
62 Himalayan events. Pre-Himalayan granites are exposed in various regions in NW Pakistan (i.e. Uthla,
63 Mansehra and Swat, Fig. 1). New U-Pb zircon ages of different units from the Uthla and Mansehra
64 granitoids are presented here and these confirm the broader petro-tectonic significance of these
65 rocks. The Mansehra granite previously yielded a whole rock Rb-Sr age of 516 ± 16 Ma (LeFort et al.,
66 1980) which is also considered to be the age of Swat granite based on their similar field and
67 petrological character (Anczkiewicz et al., 1998; DiPietro and Isachsen, 2001). These granites are

68 associated with the early Paleozoic tectonic activity reported from the northern Indian plate on
69 regional scale (LeFort et al., 1980; Pogue et al., 1992a; DiPietro and Isachsen, 2001; Cawood et al.,
70 2007). An investigation of the petrogenetic history of these granitoids, their precise radiometric
71 dating and regional tectonic correlations provide important information for unravelling the tectono-
72 magmatic evolution of northern margin of the Indian plate during the early Paleozoic.

73 **2. Regional Geology**

74 The Indian plate, Kohistan Island Arc (KIA) and Eurasian plate are the three principal tectonic
75 domains present in north-western Pakistan. The KIA originated as a result of intra-oceanic
76 subduction within the Tethys Ocean which collided with the Indian plate along a regional fault zone
77 known as Main Mantle Thrust (MMT) (Fig. 1) (Searle and Treloar, 2010; Burg, 2011). The northern
78 Indian plate is divided into three tectonic units known as the Lesser Himalayan sequence (LHS), the
79 Greater Himalayan Sequence (GHS) and the Tethyan Himalayan sequence (THS) which are separated
80 by two regional fault systems: the Main Central Thrust (MCT) and South Tibetan Detachment system
81 (STDS) (Fig. 1). Different granitic suites have been identified in Himalayan terrane i.e. early Cenozoic
82 linked to Himalayan collision (e.g. Searle et al., 2009; King et al., 2011), rift related late Paleozoic
83 (e.g. Noble et al., 2001; Ahmed et al., 2013) and early Paleozoic due to accretionary orogenesis (e.g.
84 LeFort et al., 1986; Schelling, 1999; Cawood et al., 2007; Wang et al., 2013). The early Paleozoic
85 granitic rocks intrude different GHS and LHS lithologies as confirmed by several geochronological
86 investigations (e.g. Girard and Bussy, 1999; Miller et al., 2001; Gehrels et al., 2006a).

87 The study area in this paper is located in the Peshawar basin, which lies south of the MMT within the
88 northern portion of the Indian plate (Fig. 1), today forming a portion of the Himalayan zone. The
89 Precambrian and early Paleozoic stratigraphy of the study region is summarized in Figure 2. The
90 Precambrian Tanawal formation is unconformably overlain by Cambrian Ambar formation
91 (equivalent of Abbottabad formation in Hazara region). The stratigraphic contact is marked by an
92 angular pebble conglomerate containing quartzite clasts from the Precambrian Tanawal formation

93 (Pogue et al., 1992b; 1999), which contains quartzites and meta-pelites and served as the country
94 rocks for the emplacement of the Utlā and Mansehra granites (Fig. 1). The Swat granite is hosted by
95 Manglaur formation, which is considered as the lateral correlative of the Tanawal formation
96 (DiPietro et al., 1993). The current investigation is largely based on the samples from the Utlā
97 granites, which we compared with our new as well as the already published data from the Mansehra
98 granites (Le Fort et al., 1980; DiPietro and Isachsen, 2001).

99 Our study region is mainly affected by regional Barrovian metamorphism that resulted from the
100 Tertiary collision of the Indian plate with KIA (Treloar, 1997). The systematic study of
101 metasedimentary units indicates an increase in metamorphic grade from chlorite to sillimanite grade
102 towards the collision boundary (Treloar et al., 1989a). Pre-Himalayan deformation along with
103 tectonism in the early Paleozoic have been identified in the region as shown by the metamorphic
104 overprint of Proterozoic strata (i.e. Hazara and Manglaur formations, Baig et al., 1988; Williams et
105 al., 1988) and unmetamorphosed Paleozoic sequence (Pogue et al., 1992b). Analogous pre-
106 Himalayan tectonism has also been recorded in other Himalayan regions from NW India and Nepal
107 (Gehrels et al., 2006a, b).

108 **3. Field features**

109 A sharp intrusive contact of the Utlā and Mansehra granites with Precambrian Tanawal formation
110 has been observed (Fig. 3A). On the basis of the field features, the Utlā granites can be divided into
111 mega-porphyritic granites (MPG), aplite dykes having granitic compositions with fine to medium
112 grained texture (AMG) and quartz rich veins (QRV) (Fig. 3B-D). The Mansehra granites exhibit similar
113 field characteristics to Utlā granites (MPG). The mineralogical composition and field characteristics
114 of the Utlā intrusive units are summarized in Table 1. QRV contain < 5% feldspars and range in
115 thickness from a few up to several centimetres. Apart from the textural differences, the dark
116 minerals content is less in the AMG than MPG units. Dykes of a basic composition also cross-cut the
117 host granites. Most of the granites are fresh and compact, however, foliation can be found in some

118 areas and this is particularly so adjacent to shear zones. In some places the surfaces of granitoids
119 show intense weathering leading to a dark brown colour. Tourmaline rich veins cross cut both the
120 MPG and AMG (Fig. 3E) and tourmaline nodules have also been observed in certain areas (Fig. 3F).

121 **4. Laboratory Methods**

122 **4.1. Zircon Separation**

123 We generated zircon concentrates by crushing, grinding and sieving of hand samples followed by
124 magnetic and heavy liquid separation. The samples were initially crushed and ground using a jaw
125 crusher and grinder. The resulting material was passed through 350 μm sieve to remove the coarser
126 fragments. A Wilfley table was used to separate the heavy minerals from the less dense phases. The
127 heavy mineral fractions were processed through a Frantz magnetic separator, which eliminated the
128 magnetic fraction. The remaining fractions were passed through heavy liquid (Di-iodomethane, $\rho =$
129 3.32 g/cm^3) to obtain highly concentrated zircon separates. Around 100 zircon grains from each
130 sample were embedded in a 1-inch diameter, round epoxy block, after which the individual grains
131 were imaged in both back scattered electron (BSE) and Cathode Luminescence (CL) modes using a
132 scanning electron microscope. These images revealed the internal growth zoning of the many grains,
133 allowing a better targeted selection of spots for U-Pb isotope determination.

134 **4.2. U-Pb isotopes analysis**

135 U-Pb isotopes analyses were conducted by Secondary Ion Mass Spectrometry (SIMS) using the
136 CAMECA 1280-HR instrument at GFZ Potsdam in Germany. The analyses employed a 10 nA, $^{16}\text{O}^{2-}$
137 shaped primary ion beam resulting in a beam diameter of 25 μm at the sample's surface. Positive
138 secondary ions were extracted using a +10 kV potential as applied to the sample holder. Each
139 analysis was preceded by a 10 nA, pre-sputtering lasting 120 seconds and employed a 25 μm raster.
140 Oxygen flooding was used to enhance the lead sensitivity. The data were acquired using an ETP
141 electron multiplier operating in mono-collection mode. A single analysis consisted of 16 cycles of the

142 peak stepping sequence: $^{90}\text{Zr}_2\ ^{16}\text{O}$ (1 second integration time per cycle), $^{92}\text{Zr}_2\ ^{16}\text{O}$ (1s), 200.5 (4s), $^{94}\text{Zr}_2$
143 ^{16}O (1s), ^{204}Pb (6s), ^{206}Pb (4s), ^{207}Pb (6s), ^{208}Pb (2s), $^{177}\text{Hf}\ ^{16}\text{O}_2$ (1s), ^{232}Th (2s), ^{238}U (2s), $^{232}\text{Th}\ ^{16}\text{O}$ (2s),
144 $^{238}\text{U}\ ^{16}\text{O}$ (2s) and $^{238}\text{U}\ ^{16}\text{O}_2$ (2s). Thus, including pre-sputtering, a single analysis lasted approximately
145 16 minutes.

146 The software package Isoplot (Ludwig, 2012), was used to display the data using the decay constants
147 recommended by the IUGS sub-commission on geochronology (Steiger and Jäger, 1977). Our
148 common lead correction was based on the observed ^{204}Pb counts in conjunction with the modern
149 common lead compositions from the model of Stacey and Kramers (1975).

150 The reference material (RM) zircon 91500 ($^{206}\text{Pb}/^{238}\text{U}$ age: 1062.4 ± 0.4 Ma; $^{207}\text{Pb}/^{206}\text{Pb}$ age: $1065.4 \pm$
151 0.3 , Wiedenbeck et al., 1995) was used both as primary U-Pb calibration material and to establish
152 the repeatability of the analytical method. The RM zircon Temora 2 ($^{206}\text{Pb}/^{238}\text{U}$ age: 416.78 ± 0.33
153 Ma, Black et al., 2004) was used as a quality control to evaluate data accuracy. The analysis of one or
154 two RMs was regularly interspersed after every four unknown points during data collection. The
155 analyses were carried out in two sessions (Appendix-1). In the first session, eleven determinations of
156 RM 91500 yielded a Concordia age of 1063.6 ± 4.8 Ma (2s) and a $^{206}\text{Pb}/^{238}\text{U}$ age of 1063.2 ± 14.4 Ma
157 (2s), respectively. The respective ages obtained during the second analytical session ($n = 16$) were
158 1063.4 ± 2.5 Ma (2s) and 1062.9 ± 8.4 Ma (2s). Figure 4a presents the collective concordia age
159 distribution of RM 91500 for both sessions (1063.4 ± 2.3 Ma); the respective $^{206}\text{Pb}/^{238}\text{U}$ age is 1063.0
160 ± 10.8 Ma (2s). On the basis of the 91500 results the repeatability of our concordia and $^{206}\text{Pb}/^{238}\text{U}$
161 age determinations are estimated to be better than 0.5 % and 1.5%, respectively.

162 Ten analyses of zircon RM Temora 2 obtained during the two sessions yielded a mean concordia age
163 of 414.3 ± 2.6 Ma at the 95% confidence level (Fig. 4b) and a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 413.1 ± 8.5 Ma
164 (2s). These ages are around 0.6% and 0.9% lower, respectively, than the assigned values reported by
165 Black et al. (2004). This slight offset may reflect instrumental bias and/or sample heterogeneity.
166 Overall, the repeatability of the U-Pb calibration on 91500 zircon as well as results of Temora 2
167 suggest that the data are reliable to better than $\pm 2\%$.

168

169 **4.3. Mineral Chemistry**

170 The chemical composition of minerals from representative rock samples were determined using a
171 JEOL JXA-8200 Superprobe at Camborne School of Mines (CSM), University of Exeter, UK. An
172 accelerating voltage of 20 kV, beam current 10 nA and beam diameter of 5 µm were used for
173 analysis of all mineral phases. The instrument was calibrated (elements and X-ray lines shown in
174 parentheses) using orthoclase (SiKα, KKα), almandine (FeKα, AlKα), albite (NaKα), wollastonite
175 (CaKα), periclase (MgKα), rhodonite (MnKα), rutile (TiKα), tugtupite (ClKα) and fluorite (FKα)
176 calibrants. Mineral formulas were calculated on the basis of stoichiometry and charge balance.

177 **4.4. Whole rock geochemistry (Major and trace elements)**

178 Whole rock major element compositions were determined using the Bruker S4 Pioneer wavelength
179 dispersive X-ray fluorescence (XRF) spectrometer at CSM on fused disks. Reference materials (SARM-
180 1 and DNC-1) were run at the start and at the end of each batch to monitor both drift and the
181 stability of the instrument during data acquisition. For sample preparation, fresh rock samples were
182 crushed and ground using an agate mill. The relative analytical uncertainty on the total
183 concentration of each element is <1% for SiO₂, Al₂O₃, Na₂O and K₂O, <2% for CaO, Fe₂O₃ and TiO₂ and
184 <10% for MgO, MnO and P₂O₅.

185 Trace elements, including rare earth elements (REEs) were determined by Inductively Coupled
186 Plasma Mass Spectrometry (ICP-MS) at CSM. Four acids including HF, HNO₃, HClO₄ and HCl were
187 used to digest the rock powder in Teflon tubes for analysis following the procedure of Yu et al.,
188 (2001). Two blanks and two reference solutions (Sarm-1: granite and DNC-1; Dolerite) were
189 prepared and analysed during each analytical run to assess the elemental contamination and
190 precision. A reference solution, with a known concentration of elements, was also analysed after

191 each six unknowns to observe instrumental drift. The precision is <5 % for all the element except for
192 La, Ce and Zr which is <10%.

193 **5. Results**

194 **5.1. Petrography and mineral chemistry**

195 MPG constitutes the major portion of Utlá granites, having both plagioclase and alkali feldspar
196 (including both orthoclase and microcline) as phenocrysts. Plagioclase is mostly cloudy, displaying
197 partial to complete alteration to sericite and clay minerals. It occurs as phenocrysts as well as
198 forming part of the MPG groundmass. Some of the phenocrysts display distinct zoning (Fig. 5a). The
199 albite content of zoned plagioclase grains systematically increases from core to margin (Core: An₃₆,
200 Margin: An₁₆) (Appendix-2) elaborating the fractional crystallization of the melt. Alteration to clay
201 minerals and sericite in the core domain (Fig. 5a) reflects higher Ca content as calcic plagioclase is
202 more susceptible to alteration than sodic. Plagioclase in the ground mass is mostly uniform in
203 composition ranging from An₂₃ to An₁₁ (Fig. 5b). In comparison, plagioclase from AMG is mostly
204 unzoned and has a high albite content (An_{0.22}-An₆) (Fig. 5b). Alkali feldspar, including both orthoclase
205 and microcline, occur as phenocryst as well as groundmass in the MPG. Perthite exsolution is
206 observed in orthoclase grains, suggesting their crystallization under sub-solidus conditions. Feldspars
207 from the AMG samples do not show exsolution, implying a faster cooling rate as compared to MPG.

208 Micas (both biotite and muscovite) and tourmaline are commonly occurring accessory minerals as
209 well as lesser amount of rutile, garnet, apatite and andalusite. Rounded to sub-rounded nodules
210 formed by accretion of dark coloured tourmaline is observed in AMG. In addition to quartz and
211 minor amounts of feldspar, QRV also contain well developed coarse grained (up to 1cm long) dark
212 tourmaline grains.

213 The chemical variations in the micas provide an excellent opportunity for distinguishing between the
214 various phases of magmatism. The Fe+Mn+Ti–Al^{VI} vs Mg–Li plot (Tischendorf et al. 2001) groups the
215 analyses into two clusters and thus effectively separates MPG and AMG biotites (Fig. 5c). Magnesian
216 siderophyllite (Fe biotite) dominates in the MPG samples; in contrast the composition changes
217 towards siderophyllite in the AMG. The concentration of F is relatively higher in the MPG than in the
218 AMG biotites (Appendix-2). The analyses of muscovite from both the rock types fall mostly in the
219 region of muscovite, however some of the MPG data points approach the ferroan muscovite field
220 (Fig. 5c).

221 **5.2. U-Pb Zircon Geochronology**

222 Zircons separated from the Mansehra (M-5) and Ulla granites (UT-69) and the cross cutting aplite
223 dyke (UT-44) show a large range in morphology and size (Fig. 6). Most zircons occur as subhedral,
224 short-prismatic to oval, < 20-120 µm long crystals. Cracks, holes, mineral inclusions are common,
225 limiting considerably the number of possible SIMS analytical spots. Most of zircons have old and
226 partly metamict cores surrounded by young rims which we believe to represent the crystallization
227 age of the granitic hosts. Fig. 6 show the CL images of representative zircons with position of the
228 analytic spots. Appendix-1 enlists the zircon U-Pb ages of all the samples obtained by SIMS.

229 **5.2.1. Ulla Granite (MPG, UT-69)**

230 ²⁰⁶Pb/²³⁸U ages of MPG zircons range from 894 to 458 Ma (Appendix-1). At least two major
231 crystallization events are recognizable. Proterozoic crystallization at ≈ 1100-800 Ma is documented
232 by inherited cores (Fig. 6). In contrast, a young event at around 485 to 470 Ma is documented in the
233 zircon rims, thus documenting the crystallization of the Ulla granite. The average concordia age of all
234 rim analyses younger than 500 Ma is 478.8 ± 2.3 Ma (2s; Fig. 4c). ²⁰⁶Pb/²³⁸U and ²⁰⁷Pb/²³⁵U data
235 younger than 480 Ma average at 473.2 ± 8.3 Ma (1s, n=7) and 474.2 ± 4.5 Ma (1s, n=8), respectively.

236 The crystallization of MPG zircons appear to have lasted a protracted time frame having ended at
237 around 474 Ma.

238

239 **5.2.2. Utle Granite (AMG, UT-44)**

240 The age spectra of granite UT-69 and aplite dyke UT-44 overlap, thus suggesting a common genesis
241 mechanism. Again, inherited cores are Proterozoic in age (with a single sample even being as old as
242 1700 Ma). Concordia ages of the rim domains < 500 Ma suggest a slightly younger crystallization age
243 of dyke UT-44 as compared to cross cut granite UT-69, averaging at 476.6 ± 4.1 Ma (2s; Fig. 4d). A
244 roughly 2 Ma younger age is also indicated by the $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ data younger than 480
245 Ma, which average at 471.2 ± 2.3 Ma (1s, n=5) and 472.9 ± 5.8 Ma (1s, n=5), respectively. Note
246 however that within 2s uncertainties, the ages of MPG and AMG rocks overlap and are
247 indistinguishable.

248

249 **5.2.2. Mansehra Granite (M-5)**

250 Most inherited cores in M-5 zircons range between around 600 and 1315 Ma in age, with a
251 single concordant sample being as old as 2000 Ma (Appendix-1). In marked contrast to the Utle
252 rocks, rim domains of Mansehra zircons form reverse discordant correlation lines in both the
253 conventional and the Tera-Wasserburg Concordia diagrams (Fig. 4e-f). Strongly reverse discordant
254 $^{206}\text{Pb}/^{238}\text{U}$ zircon ages have also been documented for 2700 Ma old zircons from the Murchison
255 Province of Western Australia (Wiedenbeck 1995) and for 180 Ma old zircons from the Tasmanian
256 dolerites (South-East Australia; Williams and Hergt, 2000; White and Ireland, 2012). In both cases,
257 the zircons are characterized by high uranium contents of up to 3400 $\mu\text{g}/\text{g}$ (Wiedenbeck, 1995) and
258 12000 $\mu\text{g}/\text{g}$ (White and Ireland, 2012) and display correlations between U concentration and
259 apparent $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ages. Wiedenbeck (1995) explains this effect by exclusively
260 natural processes involving mobilization and redistribution of labile Pb, while White and Ireland

261 (2012) argue for instrumental (tuning) and analytical causes (matrix effect). Rims of M-5 zircons are
262 significantly higher in uranium (1300 to 9600 $\mu\text{g/g}$) as compared to UT-69 and UT-44 zircons (< 1000
263 $\mu\text{g/g}$) and a rough correlation between U content and apparent Pb/U ages is observed (not shown).
264 We therefore hesitate to use these Pb/U data to derive a crystallization age for the Mansehra
265 granite. Instead, we suggest that the Discordia-Concordia intercepts at 479.5 ± 5.8 Ma (Fig. 4e and
266 4f) date the formation of the M-5 rock, which is statistically indistinguishable from those of the Utlā
267 granites.

268 **5.3. Whole rock major and trace elements geochemistry**

269 The concentration of major and trace elements in representative whole rock samples is presented in
270 Table 2 (complete data set is available Appendix-3). Generally, samples from both the MPG and AMG
271 are strongly per-aluminous as suggested by high A/CNK values (> 1.11 ; Fig. 7), which is also
272 supported by the mineralogical characteristics described earlier. Likely, all the samples are
273 predominantly sub-alkaline as described in total alkalis vs. silica plot (Fig. 7). Samples from Mansehra
274 granite were also analysed during the current investigation (Appendix-3) and their data is presented
275 (Fig. 7). They show similar petrographic and geochemical behaviour to that seen for the Utlā granites
276 (MPG) i.e. megacrystic, strongly peraluminous, sub-alkaline. The variation plots of major elements
277 are presented in Fig. 8. A decreasing trend in Al_2O_3 , MgO, Fe_2O_3 and CaO against SiO_2 points to
278 fractionation of plagioclase and biotite. A negative correlation between P_2O_5 and CaO with SiO_2
279 indicates the fractionation of apatite. Inclusions of apatite in biotite further support this observation.
280 TiO_2 shows a negative relationship with increasing SiO_2 content.

281 A curve trend has been observed in the plot of Rb/Sr (differentiation index) against Fe_2O_3 ,
282 mainly due to an enrichment of Rb in AMG (Fig. 8). Both rock types show LREE enriched patterns
283 with negative Eu anomalies (Fig. 9). However, the total REE content of MPG ($\Sigma\text{REE} = 89.2 \mu\text{g/g}$) is
284 greater than AMG ($\Sigma\text{REE} = 36.2 \mu\text{g/g}$) (Appendix-3). The Eu anomaly is much stronger in the case of
285 AMG ($\text{Eu}/\text{Eu}^* = 0.23$) than is seen in MPG ($\text{Eu}/\text{Eu}^* = 0.42$) (Fig. 9). The trace elements normalized to

286 primitive mantle reveal a similar pattern for both MPG and AMG (Fig. 9), showing an enrichment of
287 Rb, Th, U and Pb and a depletion of Ba, Sr, Ti, Zr and Nb.

288 **6. Discussion**

289 **6.1. Fractional crystallization and Source rock characteristics**

290 The evolution of the Uta granitoids via fractional crystallization is visible on Harker plots (Fig. 8) and
291 is also suggested by the observed mineralogical compositions. The fractionated phases mainly
292 include feldspars (plagioclase and orthoclase), biotite, apatite and zircon. A normal zoning (i.e.
293 increasing Ab content from core to rim) observed in plagioclase from MPG and an enrichment in the
294 albite content of plagioclase in AMG relative to MPG strongly indicates plagioclase fractionation.
295 Furthermore, AMG samples plot towards the higher silica content relative to MPG, implying a more
296 evolved character (Fig. 8). Despite such a strong evidence for fractionation process, the total REE
297 content of AMG is less than MPG. This can be explained by the fractionation of REE bearing phases in
298 MPG, including zircon and apatite, which are most likely hosts for REEs. The presence of these
299 minerals as inclusions in biotite supports this interpretation.

300 The negative Eu anomaly in the chondrite normalized REE pattern also supports the
301 fractionation of plagioclase as Eu^{2+} normally substitute for Ca in feldspar lattice (Rollinson, 1993).
302 The observed negative Eu anomalies could also reflect the fractionation of other minerals including
303 hornblende, titanite, garnet; however this would result in the opposite trend to that of feldspar. The
304 observed higher LREE/HREE ratio is very typical of granitic systems and is commonly caused by the
305 fractionation of zircon from the magma (Rollinson, 1993).

306 The negative Ba, Sr, Nb and Ti anomalies (Fig. 9) also points to a highly fractionated nature of
307 the studied granites (Chappell, 1999). Fractionation of Ti bearing phases (e.g. rutile, ilmenite and
308 sphene) normally results in a negative Ti anomaly. Negative Sr and Ba anomalies are indicative of
309 plagioclase and orthoclase fractionation, respectively (e.g., Miller, 1985; Patiño Douce et al., 1990;
310 Wu et al., 2003; Healy et al., 2004; Li et al., 2015). The identical geochemical behaviour of Uta and

311 Mansehra granitoids (Fig. 7 and 9) show their origin related to a single coeval magmatism (Sajid et
312 al., 2014).

313 Various experiments have been performed for the estimation of source rock features of
314 peraluminous granites (e.g. Patino Douce and Johnston, 1991; Chappell and White, 1992; Skjerlie
315 and Johnston, 1996; Sylvester, 1998). The source characteristics of studied granites have been
316 interpreted using the plots of experimental data (based on Rb/Sr, Rb/Ba, CaO/Na₂O, FeO, MgO,
317 Al₂O₃, TiO₂) gathered in these investigations. The Al₂O₃/TiO₂ and CaO/Na₂O ratios from the Utlā
318 (MPG and AMG) and Mansehra granites point to a pelitic source (clay-rich, plagioclase-poor) for
319 their parental melt (Fig. 10). The molar Al₂O₃/(MgO + FeO_T) vs. CaO/(MgO + FeO_T) of Altherr et al.
320 (2000) diagram also suggests a similar source rock composition for these granitoids (Fig. 10).

321 **6.2. Comparison with other granitic suites**

322 Granites with analogous geochemical signatures in similar tectonic regimes exist in other Himalayan
323 regions (e.g. LeFort et al., 1986). In NW India several peraluminous granitic plutons with similar
324 mineralogical and geochemical characteristics were reported (e.g. Jispa granites; Islam and
325 Gururajan, 1997), Tso Morari granite (Girard and Bussy, 1999), Rakcham granite (Kwatra et al., 1999)
326 and Mandi granites (Miller et al., 2001). Similar granitic suites from Nepal include Simchar granite
327 (LeFort et al., 1983), Palung granite (Gehrels et al., 2006a) and Kathmandu (Gehrels et al., 2006b).
328 Paleogeographic reconstructions indicate a juxtapositioning of the southern China blocks (e.g.
329 Baoshan block) to the northern Indian plate in early Paleozoic time (Metcalf, 1996, Fig. 11). The
330 granitic plutons in SW China also show similar geochemical characteristics (Wang et al., 2010, 2013;
331 Dong et al., 2013) to the Utlā and Mansehra granites as well as to other Himalayan granites
332 described above (Fig. 7). The origin of these granites has been related to a melting of a
333 metasedimentary source rock (Fig. 10) followed by the fractional crystallization of feldspar and
334 biotite (Dong et al., 2013; Wang et al., 2013). The whole rock geochemical data from the Tso-Morari
335 and Mandi pluton from NW India (Girard and Bussy, 1999; Miller et al., 2001, respectively) and from

336 the Baoshan and Shan-Thai block (Fig. 11) from SW China (Dong et al., 2013; Wang et al., 2013) are
337 presented for comparison (Fig. 7 and 10). Apart from geochemical similarities, these granites yield
338 early Paleozoic emplacement ages ranging from ~460 to ~530Ma (Fig. 11) as documented by
339 extensive geochronological Rb-Sr (e.g. LeFort et al., 1980; Debon et al., 1981; Trivedi, 1990; Einfalt et
340 al., 1993) and U-Pb (e.g. Gehrels et al., 2006a; Dong et al., 2013; Wang et al., 2013) studies. The
341 geochemical and geochronological similarities of Utlā and Mansehra granites, as well as other
342 granites from the Himalayan terrane along the northern margin of Gondwana suggest an extensive
343 orogenic thermal activity during early Paleozoic (Girard and Bussy, 1999; Miller et al., 2001; Cawood
344 et al., 2007).

345 **6.3. Tectono-magmatic Implications**

346 The Himalayan region of NW Pakistan was part of the northern Gondwana supercontinent during
347 late Precambrian and Paleozoic times, constituting the Indian passive continental margin prior to
348 collision with Eurasian continent in early Cenozoic (Brookfield, 1993; Cawood et al., 2007; Wang et
349 al., 2013). The large volume of Cambro-Ordovician (~530-470Ma, Fig. 11) granitoids intruding in LHS
350 and GHS (DeCelles et al., 1998, 2000, 2004; Godin et al., 2001; Gehrels et al., 2003; Booth et al.,
351 2004; Lee and Whitehouse, 2007; Quigley et al., 2008) indicate an early Paleozoic orogenic event in
352 the Himalayan terrane.

353 **6.3.1. Evidence of early Paleozoic orogenesis**

354 The existence of this orogeny is supported by a break in the stratigraphic record (Garzanti et al.,
355 1986; Le Fort et al., 1994; Valdiya, 1995; Liu et al., 2002; Zhou et al., 2004; Qasim et al., 2017), by
356 detrital zircon age distributions from sedimentary strata (Hodges, 2000; Kusky et al., 2003; Zhang et
357 al., 2008a, b; Dong et al., 2009; Myrow et al., 2010; Spencer et al., 2012) and coeval metamorphism
358 (e.g. Argles et al., 1999; Catlos et al., 2000, 2002; Godin et al., 2001; Gehrels et al., 2003, 2006a,
359 2006b; Kohn et al., 2004) across wide regions of the Himalayas.

360 The unconformable contact marked by conglomeratic beds between the Cambro-Ordovician strata
361 with older rocks in Himalayan terrane provide strong evidence for a major period of erosion and
362 uplift associated with an orogenic event during the early Paleozoic (e.g. Garzanti et al., 1986;
363 Brookfield, 1993; Qasim et al., 2017). An important observation is that lower Ordovician basal
364 conglomerates unconformably overly the Cambrian strata in Kathmandu, Nepal (Kumar et al., 1978;
365 Funakawa, 2001; Gehrels et al., 2003). Similarly, in the Spiti and Zaskar regions of NW India, the
366 Cambrian lithologies are unconformably overlain by coarse Ordovician conglomerates (Hughes,
367 2002; Myrow et al., 2006a, b). The U-Pb chronology and Hf isotopes of detrital zircons from
368 sedimentary strata led Spencer et al. (2011, 2012) to propose that sediment input in GHS was
369 prompted by orogenic activity in the early Paleozoic. A similar interpretation was also made through
370 the comparison of U-Pb age of detrital zircons representing a particular time span from the broad
371 geographic range covering around 2000 km along the length of Himalayan orogen from Pakistan to
372 Bhutan (Myrow et al., 2010). Evidence for early Paleozoic deformation and metamorphism include
373 regional folding and metamorphism (up to garnet grade) of Neoproterozoic Bhimphedi group rocks
374 prior to the deposition of overlying Ordovician strata (Gehrels et al., 2006a). All these observation
375 points to an early Paleozoic orogenic phase in the northern Gondwana that extended at least from
376 NW India to the Namche Barwa region in southern China (Wang et al., 2013).

377 The early Paleozoic stratigraphy of eastern Peshawar basin in NW Pakistan is summarized in Fig. 2.
378 The Precambrian Tanawal formation is unconformably overlain by Cambrian Ambar formation. The
379 contact is marked by angular pebble conglomerate containing quartzite clasts from the Tanawal
380 formation (Pogue et al., 1992b; 1999). In certain regions (i.e. northern Swabi close to Utla area) the
381 Ambar formation is completely eroded and Tanawal formation is directly overlain by the Ordovician
382 Misri Banda quartzites (Pogue et al., 1999). The Abbottabad formation is the lateral equivalent of
383 Ambar formation in Hazara region east towards the Indus river (Fig. 1). The unconformable contact
384 between Abbottabad and Tanawal formation is noticed, however, in certain regions, Tanawal
385 formation is eroded and Abbottabad formation directly overlain the Precambrian Hazara formation

386 through an angular unconformity marked by conglomeratic bed containing angular clasts (Calkins et
387 al., 1975; Baig et al., 1988; Pogue et al., 1999; Palin et al., 2018). The growth of new metamorphic
388 mica along with the presence of a pronounced cleavage has been noticed in the Hazara formation,
389 which has been metamorphosed up to the greenschist facies (Baig et al., 1988). Metamorphism of
390 Tanawal formation has also been reported in the recently investigated U-Pb monazite
391 geochronology (482.4 ± 7.9 Ma and 464.5 ± 4.0 Ma, Palin et al., 2018). No evidence of deformation
392 and metamorphism is observed in either the conglomeratic bed marking the unconformity nor in the
393 overlying Abbottabad formation. These indications imply a tectonothermal event during the early
394 Paleozoic times, which deformed and metamorphosed the Precambrian strata. The intrusion of
395 peraluminous Utlā and Mansehra granites (478.8 ± 2.3 Ma and 479.5 ± 5.8 Ma respectively) in this
396 region is consistent with the existence of such an event.

397 **6.3.2. Tectonic model**

398 Our U-Pb zircon dating documents the crystallization of the peraluminous Utlā and Mansehra
399 granites between 470 and 480Ma (Fig. 4 and 6). Similar aged rocks are also present in other
400 Himalayan regions, including both NW India (e.g. Miller et al., 2001) and Nepal (e.g. Gehrels et al.,
401 2006a). Based on the tectonic position, age, geochemistry of studied granites and the earlier
402 mentioned evidences of orogenesis, we propose a hypothetical tectonic model. It elaborates the
403 origination of granites from the melting of metasedimentary rocks due to the large scale thermal
404 activity triggered by the collision of the northern margin of Gondwana with southward subducted
405 oceanic lithosphere of the proto-Tethys (Fig. 12). The outlines of the events proposed are described
406 as under:

- 407 1. The integration of current geochemical and geochronological data with the existing data set
408 indicates a convergent margin setting of northern Indian plate in the early Paleozoic
409 (Cawood et al., 2007). At that time the Prototethyan oceanic crust was subducted beneath
410 northern Gondwana, which produced the extended back-arc basin due to roll-back of the

411 subducted plate (Wang et al., 2013), leading to sediment accumulation in a back-arc basin
412 setting (Palin et al., 2018). Pogue et al. (1992b) and Qasim et al. (2017) presented the
413 regional scale Proterozoic and early Paleozoic stratigraphy of Hazara and Swabi region, north
414 Pakistan. The constituent lithologies mainly include argillites, quartzites and carbonates
415 which shows the existence of back-arc basin and subsequent sediment build-up (Palin et al.,
416 2018).

417 2. The existence of magmatism along the traverse of Himalayan orogen (around 700km,
418 Cawood et al., 2007) requires large scale crustal anatexis, which could not have been
419 induced by a simple magmatic arc. This suggests the crustal melting in the extensive back-arc
420 region induced by the high heat flow. The evidence of coeval mafic magmatism has been
421 highlighted by earlier workers (e.g. Miller et al., 2001). The Mandi mafic rocks in GHS of NW
422 India yielded 496 ± 14 Ma (Sm-Nd isotopes) age and exhibit convergent margin geochemical
423 signature (Miller et al., 2001). The cumulated gabbros (473 ± 3.8 Ma) from Baoshan block
424 (Wang et al., 2012) and associated synchronous I-type granites are believed to be the
425 product of the subduction of the Prototethyan Ocean (e.g. Zhang et al., 2008; Liu et al.,
426 2009; Zhu et al., 2012). The similar aged basic magmatism in the region suggests a heat input
427 from a mafic magma underplating the lower crust; such a heat source might have been
428 responsible for the large scale anatexis.

429 3. The orogenesis was related to the ongoing plate subduction followed by the accretion of an
430 isolated block onto the continental margin (Fig. 12). Different terranes or geological blocks
431 including Lhasa, Qiantang and Helmand along the Indian plate margin are recognized as part
432 of the reconstruction of northern Gondwana (Fig. 11) in the early Paleozoic (Sengor et al.,
433 1988; Metcalfe, 1996). Detrital zircon ages of early Paleozoic strata from the western
434 Qiantang also indicate its juxtaposition to the Himalayan terrane in the early Paleozoic (Zhu
435 et al., 2013). The vicinity of these block to the Indian plate is also supported by paleoclimatic
436 and palaeontological records presented in Metcalfe (1996).

437 4. The peraluminous signature of most of granitic plutons in the northern Indian plate (e.g.
438 Girard and Bussy, 1999; Miller et al., 2001; Wang et al., 2013; current study) support the
439 accretion of one or more outboard blocks with the continental margin (Fig. 12), which led to
440 significant crustal and sediment thickening in the back-arc basin, which along with the mafic
441 underplating, would have been responsible for crustal melting and generation of granitic
442 melts. The prime factor for generating the peraluminous S-type granite is the melting of
443 meta-sedimentary source, most likely due to deep burial caused by continent-continent
444 collision (Harris et al., 1986; Tingyu et al., 1995). The collision of an outboard block with the
445 continental margin provided a similar environment that led to the formation of
446 peraluminous melts (Fig. 12).

447 5. The partial melting and roll back of subducted oceanic lithosphere and the subsequent slab
448 break-off also induced mantle convection, which result in the formation of coeval calc-
449 alkaline volcanic rocks (Miller et al., 2001; Visona et al., 2010). The involvement of mantle in
450 the orogenic process is also shown by the presence of cumulate gabbros in the Qiangtang
451 (467 Ma) and Baoshan (473 ± 3.8 Ma) blocks (Zhai et al., 2009; Wang et al., 2012
452 respectively). Cambrian volcanic rocks include volcanic tuffs, basalts, andesites and felsic
453 volcanic rocks in the western Tethyan Himalaya (e.g., Brookfield, 1993; Garzanti et al., 1986;
454 Valdiya, 1995). The occurrence of an extensive subduction zone magmatism along the Indian
455 margin of Gondwana is shown due to the presence of basalts and basaltic andesites from
456 different regions of GHS including Mandi pluton, western Himalayas (Miller et al., 2001) and
457 Kharta region, central-eastern Himalayas (Visona et al., 2010). Late Cambrian (~ 492 Ma)
458 bimodal volcanic rocks have been identified in the central Lhasa block which relate to an
459 active continental margin (Ji et al., 2009; Zhu et al., 2012). Subduction related late Cambrian
460 (499.2 ± 2.1 Ma) mafic lavas have also been reported in Gongyanghe Group in Baoshan block
461 (Yang et al. 2012) overlain by lower Ordovician basal conglomerate (Wang, 2000).

462 The existing U-Pb geochronology data set reveal the emplacement of the majority of granitic
463 plutons during 470-490Ma span (e.g. Debon et al., 1981; Trivedi et al., 1986; Girard and Bussy, 1999;
464 Gehrels et al., 2006a) which marks the active span for crustal melting and granite emplacement.
465 However, relatively few ages have been reported between 500-530Ma (e.g. Lee et al., 2004, 2006),
466 which most probably was the time of onset of tectonic activity (Cawood et al., 2007). According to
467 petrological data of the studied granites in combination with the model presented here, an Andean
468 type orogeny along the northern margin of the Gondwana supercontinent took place during the
469 early Paleozoic.

470 **6.3.3. Other Tectonic models**

471 On the basis of available Rb-Sr and K-Ar ages of rock suites in Africa, Kennedy (1964) for the first
472 time introduced the concept of the Pan-African orogeny. This includes thermo-tectonic events
473 occurring around 500Ma, during which several mobile belts formed around the periphery of the
474 older cratons. This concept was extended to all the regions of Gondwana, which also include the
475 Indian plate. The western end of Indian plate was attached to present day Somalia and Kenya with
476 Madagascar in between (Smith et al., 1981; Dalziel, 1992; Unrug, 1996). The existence of analogous
477 geological features confirms the pan-African events in Madagascar and Southern India, which
478 includes the occurrence of major shear zone having >100km sinistral displacement and analogous
479 podiform chromite deposits in ultramafic lenses (Kroner and Stern, 2004).

480 However, the similar aged magmatism of the northern Indian plate and its relationship with
481 pan-African orogenesis is unclear. Girard and Bussy (1999) relate the A-type magmatism in Kaghan
482 granite (Pakistan) and Rupshu granite (Ladakh, India) with the youngest A-type activity from the
483 Arabian-Nubian shield in the Pan-African belt. This comparison seems uncertain due the subtle age
484 differences (Trivedi et al., 1986; Windley et al., 1996) in both events. Kwatra et al. (1999) described
485 the geochemical and geochronological features of early Paleozoic Rakcham granitoids from Sutlej
486 valley in NW India. Based on similar tectonic settings and similar trace element discrimination

487 diagrams, they include the Cambro-Ordovician granitic suites (including Mansehra granite) of the
488 northern Indian plate with the Pan-African magmatic belt.

489 In contrast, Stern (1994) clearly shows that the early Paleozoic granites in NW Himalayas post-
490 date the collision and major thermal events of the Pan-African belt. Only the southern part of the
491 Indian plate is affected by these events, as revealed by their paleogeographic reconstruction (Stern,
492 1994). Furthermore, in most of the Pan-African suites (e.g. Mali, southern Brazil) a general magmatic
493 evolutionary trend is observed with subduction related rocks (Bonin et al., 1998). The alkaline
494 magmatism during similar time span also forms a common observation in these belts. Although
495 having a roughly similar crystallization age to the Utlā-Mansehra granites (and other Cambro-
496 Ordovician GHS and LHS granites), the convincing evidence for their association with Pan-African
497 event is lacking due to the absence of any evolutionary imprints from mafic magma and lack of
498 synchronous alkaline magmatism.

499 The early Paleozoic magmatism was related to a non-arc extensional setting (Miller et al.,
500 2001). The granites studied here fall in the field of syn-collisional to volcanic arc settings based on Nb
501 vs Y and Rb vs Y+Nb discrimination diagrams (Fig. 13). Moreover, these granites are depleted in Ba,
502 Nb and Sr but enriched in Rb, Th and U (Fig. 9), indicative of continental volcanic arc settings (Brown
503 et al., 1984). This geochemical evidence contradicts an extensional non-arc setting in favour of a
504 magmatic arc environment, which was most probably associated with subduction and collision along
505 the northern margin of Gondwana. This interpretation is supported by the widespread existence of
506 volcanic rocks, including acid and basic tuffs and andesites in the western Tethyan Himalayas
507 (Garzanti et al., 1986; Valdiya, 1995) and Lhasa terrane (Zhu et al., 2012).

508 **6.3.4. Implications for Late Proterozoic magmatism in northern Indian plate**

509 The concordant core ages from the studied zircons (both Utlā and Mansehra Granites) mostly
510 ranges between 813 Ma to 893 Ma. Similar U-Pb zircon ages have been reported from Black

511 Mountain Complex (DiPietro and Isachsen, 2004) and Chingalai Gneisses (Ahmed et al., 2013). The
512 Black Mountain Complex is present along the eastern bank of Indus River mainly composed of
513 equigranular fine-grained biotite-quartz-feldspar orthogneisses and intruded by Mansehra Granite
514 (DiPietro and Isachsen, 2004). Other subordinate lithologies include biotite and garnet bearing
515 gneisses, migmatites and mafic rocks. Zircon from the main lithology yielded an age of 823 ± 2 Ma
516 (DiPietro and Isachsen, 2004). The Chingalai Gneisses are present on the western side of Indus River
517 and mapped as south-eastern extension of Swat Granite by DiPietro et al. (1998). The U-Pb zircon
518 age of 816 ± 70 Ma has been reported from these gneisses (Ahmed et al., 2013). Similar range of
519 these ages and ages obtained from zircon cores in current investigation points towards the
520 occurrence of small scale magmatism in late Proterozoic time in the northern Indian plate. DiPietro
521 and Isachsen (2004) and Ahmed et al. (2013) relate these rocks with Malani magmatism, which show
522 the age range between 750 and 850 Ma on the Aravalli craton exposed on the Rajasthan platform of
523 northern India.

524 **7. Conclusions**

525• The Utlā granites intrude Precambrian metasedimentary rocks of the northern Indian plate, NW
526 Pakistan. The granites are megacrystic and contain biotite, muscovite, tourmaline, garnet, epidote
527 and apatite as accessory phases. Fine to medium grained aplite dykes and coarse-grained quartz rich
528 veins cross cut the megacrystic granites.

529• The granitoids are S-type and strongly peraluminous (A/CNK values > 1.11). Fractional crystallization
530 is indicated as the dominant evolutionary process due to the change in composition of plagioclase
531 from megacrystic granite to aplite dykes, zoned plagioclase megacrysts and higher silica content in
532 the aplite dykes. Major and trace elements geochemistry imply the origin of these granites is related
533 to the melting of a pelitic source.

534• The early Paleozoic 478.8 ± 2.3 Ma emplacement age of megacrystic Utlā granites is shown by U-Pb
535 zircon geochronology followed by the intrusion of aplite dykes in 476 ± 4.1 Ma. The mineralogically
536 and geochemically similar Mansehra granites show a statistically synchronous emplacement time i.e.

537 479.5 ± 5.8 Ma.

538• The composition and geochronology of the studied plutons suggest their association with other early
539 Paleozoic granitoids from northern Gondwana. The origin of these granitoids in an arc setting related
540 to subduction of the Proto-Tethys Ocean beneath the northern Gondwana supercontinent in early
541 Paleozoic has been modelled, which is supported by the several metamorphic and stratigraphic
542 arguments as well as by the observed trace element characteristics.

543• The concordant core ages (813 Ma to 893 Ma) of zircons from both the Ubla and Mansehra granites
544 agree with published data and point towards a late Proterozoic magmatic event along northern
545 Indian plate.

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Figure Captions

907 Fig. 1. A) Geologic map of the Himalayan terrane (redrawn after Kohn 2014), star showing location of Fig. 1B,
908 B) Regional tectonic map of NW Pakistan elaborating the major granitic suites, inset showing the
909 geological map of the study area

910 Fig. 2. Precambrian and Paleozoic stratigraphy of Peshawar basin after Pogue et al., (1992b; 1999). H = Major
911 Hiatus

912 Fig. 3. Field photographs: A) Sharp contact between Utla granite and Tanawal formation, B) Feldspar
913 megacrysts in Utla granites, C) fine grained aplite in MPG D) cross cutting QRV, E) tourmaline vein in
914 Utla granite, F) nodular tourmaline in Utla granites, inset showing close-up of polished tourmaline
915 nodule

916 Fig. 4. U-Pb Concordia plots of reference materials and zircons from studied granites. The data acquired from
917 zircon rim domains of studied zircons is shown

918 Fig. 5. A) Micrograph showing zoned plagioclase phenocryst, B) Composition of plagioclase, symbols: Ab:
919 albite, Olig: Oligoclase, And: Andesine, Lab: Labradorite, Byto = bytownite, An: Anorthite, C)
920 classification of biotite and muscovite after Tischendorf et al. (2001)

921 Fig. 6. Cathode Luminescence (CL) images showing the internal structure of representative zircon grains and
922 their respective ages in Ma. Circles show the position of analytical spots

923 Fig. 7. Classification of Utla granitoids, MPG and AMG show peraluminous and sub-alkaline signature (A) $Al_2O_3/$
924 $(CaO+Na_2O+K_2O)$ versus $Al_2O_3/(Na_2O+K_2O)$ (Maniar and Piccoli, 1989) B) Alkalis vs SiO_2 plot after
925 Miyashiro (1973). The corresponding data of Mansehra granites (this study), Baoshan Block, SW China
926 (Dong et al., 2013), Shan-Thai block, SW China (Wang et al., 2013), Tso Moriri granites (Girard and
927 Bussy, 1999) and Mandi Granite (Miller et al., 2001) is also shown for comparison

928 Fig. 8. Major element harker diagrams of MPG and AMG. Symbols are same as in Fig. 7

929 Fig. 9. A) Chondrite normalized REE pattern show stronger Eu anomaly in AMG, normalized data from
930 McDonough and Sun (1995), B) Primitive mantle-normalized multi-element diagram, normalized data

931 from McDonough et al. (1992). The trend of Mansehra granite samples (blue line) analysed in this study
932 showing the average values (n = 9) is also presented along the MPG samples

933 Fig. 10. Interpretation of source rock composition A) $\text{Al}_2\text{O}_3/\text{TiO}_2$ versus $\text{CaO}/\text{Na}_2\text{O}$, field of strongly
934 peraluminous rocks from Sylvester (1998) (B) Rb/Ba versus Rb/Sr diagram, calculated source
935 composition after Sylvester, 1998, C) molar $\text{Al}_2\text{O}_3/(\text{MgO} + \text{FeO}_T)$ vs. $\text{CaO}/(\text{MgO} + \text{FeO}_T)$, source fields
936 after Altherr et al. (2000). Data source of other mentioned plutons are same as in Fig. 7.

937 Fig. 11. Early Paleozoic reconstruction map of the India–Australia prototethyan margin showing
938 paleogeographical locations of the microcontinents (revised after Cawood et al., 2007 and Wang et al.,
939 2013). The ages and location of coeval granitic pluton along the northern Gondwana margin are also
940 shown.

941 Fig. 12. Sketch showing the tectonic model of early Paleozoic orogenesis (not to scale), A) subduction of
942 Prototethyan oceanic lithosphere beneath northern margin of Indian plate. Initiation of extended back-
943 arc basin due to mantle convection and slab bending for sediment accumulation. Mafic and calc-
944 alkaline subduction related magmatism also initiate with slab bending, (B) Collision of external
945 geological block with continental margin. Crustal thickening along with mafic underplating cause partial
946 melting of lower crust to generate peraluminous melt

947 Fig. 13. Studied granites show volcanic arc to syn-collisional tectonic settings on trace elements discrimination
948 plots (after Pearce et al., 1984). VAG=volcanic arc granites, WPG= within plate granites, ORG=ocean
949 ridge granites, Syn-COLG= syn-collisional granites

Table 1. Field and mineralogical features of Utlá granites

Magmatic Unit	Field characteristics	Major Minerals	Minor assemblage (dominant minerals in bold letters)
Mega-porphyritic granite (MPG)	K-feldspar and plagioclase phenocrysts, medium to coarse grained groundmass, tourmaline and biotite recognisable in field, foliated along shear zones	Plagioclase, K- feldspar (orthoclase and microcline), quartz	Tourmaline, biotite, muscovite , andalusite, apatite, rutile, ilmenite, zircon, beryl
Fine to medium grained aplite dykes (AMG)	Texturally homogenous, nodular tourmaline in places	Plagioclase, orthoclase, quartz	Tourmaline, biotite, muscovite, epidote, garnet , rutile, zircon
Quartz rich veins (QRV)	Coarse grained, extremely fresh and white in colour	Quartz, feldspars (<5%)	Tourmaline

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Table 2. Whole rock major (wt.% age) and trace ($\mu\text{g/g}$) element concentration of representative samples from studied granites

Sample	UT-65	UT-66	UT-69	UT-40	UT-48	UT-27	UT-29	UT-30	UT-36	UT-44	M-5	M-11	M-17	M-22	M-27
Type	MPG	MPG	MPG	MPG	MPG	AMG	AMG	AMG	AMG	AMG	MG	MG	MG	MG	MG
SiO₂	71.38	68.87	71.89	75.12	71.26	76.77	77.01	74.32	76.02	76.30	70.05	72.57	69.37	69.81	71.66
TiO₂	0.61	0.49	0.35	0.24	0.69	0.14	0.14	0.12	0.10	0.11	0.56	0.21	0.66	0.53	0.46
Al₂O₃	13.97	14.04	14.73	13.03	14.29	12.49	12.82	13.87	12.96	13.34	14.86	14.47	15.00	15.15	14.17
Fe₂O₃	4.07	3.15	2.55	2.10	2.50	1.40	0.56	1.40	1.52	0.92	3.85	1.70	4.77	3.54	3.24
MnO	0.06	0.05	0.05	0.02	0.05	0.02	0.01	0.04	0.03	0.01	1.06	0.39	1.31	0.95	0.85
MgO	1.11	0.75	0.63	0.58	0.92	0.27	0.14	0.22	0.15	0.41	0.06	0.03	0.07	0.06	0.05
CaO	1.78	0.62	0.86	0.76	1.61	0.19	0.14	0.58	0.34	0.24	1.51	0.72	1.57	1.35	1.19
Na₂O	2.39	2.36	2.31	3.99	3.53	2.15	2.53	2.66	2.65	3.75	2.58	3.04	2.43	2.47	2.36
K₂O	3.55	5.30	5.15	3.22	2.87	5.33	5.97	5.58	5.42	4.11	4.15	5.28	4.12	5.02	4.31
P₂O₅	0.14	0.16	0.20	0.13	0.13	0.10	0.08	0.10	0.15	0.11	0.21	0.27	0.24	0.18	0.18
LOI	0.81	1.48	1.21	0.79	2.15	1.03	0.78	0.98	0.79	0.82	0.82	0.60	0.47	0.78	1.37
Total	99.90	97.36	99.96	100.04	100.02	99.88	100.18	99.88	100.14	100.11	99.74	99.31	100.04	99.91	99.88
Trace and REEs															
V	56.53	36.81	25.54	13.10	47.38	5.11	1.79	4.57	1.63	2.33	35.57	8.73	67.27	28.85	23.75
Cr	26.46	15.95	13.48	8.09	23.14	3.18	2.03	2.68	1.67	1.43	17.94	4.04	29.70	16.34	14.13
Rb	145.36	187.92	220.53	105.83	150.06	201.75	213.72	346.67	312.64	100.59	142.12	213.08	204.95	150.56	167.11
Sr	63.99	46.82	27.64	78.12	46.40	14.42	20.63	15.38	7.55	13.92	36.35	16.26	61.73	41.01	30.70
Y	19.85	12.89	14.53	18.60	47.53	10.25	6.99	20.74	21.71	9.40	16.10	8.61	19.89	11.79	17.70
Zr	2.98	2.56	13.91	6.21	2.44	11.18	11.07	21.57	16.33	22.93	8.14	11.74	7.95	5.47	13.68
Nb	16.02	13.53	15.39	8.70	12.49	8.46	10.33	8.55	8.54	5.81	14.52	12.97	22.89	12.52	12.47
Sn	6.74	10.71	13.59	4.41	5.47	19.84	17.63	10.20	10.16	10.54	5.41	8.96	7.06	5.35	9.71
Ba	372.39	701.82	209.83	389.49	178.11	134.46	74.78	100.85	14.91	100.49	271.88	157.35	433.17	443.72	235.13
La	35.21	17.07	17.92	16.86	72.65	6.28	3.85	11.25	6.37	5.13	23.95	15.59	58.34	23.22	20.41
Ce	65.24	30.88	25.27	26.15	126.70	12.27	6.35	22.46	11.79	7.51	46.85	34.76	127.61	46.91	46.50
Pr	8.94	4.55	4.69	4.47	17.68	1.83	1.06	3.12	1.95	1.33	6.99	4.35	15.34	6.56	5.69
Nd	33.84	17.26	17.86	16.57	65.44	6.90	3.86	11.45	7.15	4.85	27.52	16.52	58.99	25.81	22.23
Sm	7.07	3.84	4.13	3.98	13.65	1.85	1.06	3.00	2.29	1.35	5.89	4.03	12.04	5.29	4.95
Eu	1.04	0.83	0.66	0.49	1.59	0.13	0.09	0.29	0.06	0.13	0.73	0.35	1.03	0.75	0.55
Gd	5.97	3.52	3.64	3.65	11.93	1.73	1.02	2.88	2.45	1.30	4.84	3.25	9.09	4.19	4.16
Dy	4.29	2.95	3.14	3.84	10.09	2.03	1.36	3.92	3.84	1.61	3.68	2.22	5.32	2.94	3.55
Ho	0.75	0.56	0.56	0.74	1.80	0.40	0.26	0.80	0.78	0.32	0.64	0.34	0.79	0.49	0.67
Er	1.99	1.48	1.56	2.16	4.87	1.20	0.78	2.38	2.49	0.98	1.72	0.84	1.74	1.20	1.88
Tm	0.28	0.24	0.22	0.31	0.66	0.18	0.12	0.38	0.40	0.15	0.23	0.12	0.20	0.15	0.27
Yb	1.74	1.37	1.38	1.86	4.18	1.22	20.81	2.57	2.74	0.98	1.33	0.69	1.04	0.89	1.75
Lu	0.26	0.26	0.18	0.28	0.65	0.16	0.13	0.38	0.39	0.13	0.16	0.08	0.13	0.11	0.25
Hf	0.10	0.15	0.49	0.30	0.19	0.42	0.46	0.79	0.58	0.81	0.24	0.36	0.25	0.16	0.51
Ta	1.58	1.62	2.15	0.99	1.18	1.37	1.74	1.41	1.23	0.62	1.74	2.20	1.78	1.51	1.89
W	3.72	4.24	4.90	2.10	5.42	2.64	2.05	1.86	1.90	1.50	2.28	3.12	3.10	0.90	2.23
Pb	17.03	11.73	10.66	3.59	11.40	4.01	8.56	17.79	10.93	3.93	11.89	8.08	24.81	17.55	17.15
Th	15.51	8.77	8.88	8.81	16.92	6.22	4.62	10.90	8.01	3.08	16.25	12.37	30.70	14.55	14.15
U	4.52	0.86	1.71	1.09	1.88	2.07	2.64	1.78	2.97	0.56	1.46	1.32	2.94	0.93	1.94
ΣREE	166.62	84.81	81.19	81.35	331.89	36.19	20.77	64.88	42.69	25.78	124.52	83.14	291.67	118.53	112.85
A/CNK	1.27	1.31	1.34	1.14	1.21	1.29	1.18	1.21	1.19	1.21	1.29	1.20	1.32	1.27	1.32
Eu/Eu*	0.49	0.69	0.52	0.39	0.38	0.22	0.27	0.31	0.08	0.30	0.42	0.30	0.30	0.49	0.37

954 MPG and AMG = Uta granites, MG = Mansehra granites,

955 LOI = loss on ignition,

956 A/CNK = molar ratio $\text{Al}_2\text{O}_3 / (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$,

957 $\text{Eu}/\text{Eu}^* = \text{Eu}_N / (\text{Sm}_N * \text{Gd}_N)^{0.5}$