

1 **The geodynamic significance of continental UHP exhumation: New constraints from the**
2 **Tso Morari Complex, NW Himalaya**

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21 **Key Points**

- 22 1. Geochronological constraints from the Tso Morari dome place the onset of exhumation from
23 ultrahigh-pressure conditions at 46.91 ± 0.07 Ma
24 2. Close correspondence to other ultrahigh-pressure ages imply a similar time-frame for the onset of
25 UHP exhumation across the NW Himalaya
26 3. The onset of UHP exhumation at 47-46 Ma timestamps geodynamic changes in the India-Asia
27 collision zone, which triggered an Eocene plate network reorganisation

28 **Abstract**

29 The burial and exhumation of continental crust to and from ultrahigh-pressure (UHP) is an important
30 orogenic process, often interpreted with respect to the onset and/or subduction dynamics of continent-
31 continent collision. Here, we investigate the timing and significance of UHP metamorphism and
32 exhumation of the Tso Morari complex, North-West Himalaya. We present new petrochronological
33 analyses of mafic eclogites and their host-rock gneisses, combining U-Pb zircon, rutile and xenotime

34 geochronology (high-precision CA-ID-TIMS and high-spatial resolution LA-ICP-MS), garnet element
35 maps, and petrographic observations. Zircon from mafic eclogite have a CA-ID-TIMS age of $46.91 \pm$
36 0.07 Ma, with REE profiles indicative of growth at eclogite facies conditions. Those ages overlap with
37 zircon rim ages (48.9 ± 1.2 Ma, LA-ICP-MS) and xenotime ages (47.4 ± 1.4 Ma; LA-ICP-MS) from
38 the hosting Puga gneiss, which grew during breakdown of UHP garnet rims. We argue that peak zircon
39 growth at 47-46 Ma corresponds to the onset of exhumation from UHP conditions. Subsequent
40 exhumation through the rutile closure temperature, is constrained by new dates of 40.4 ± 1.7 and 36.3
41 ± 3.8 Ma (LA-ICP-MS). Overlapping ages from Kaghan imply a coeval time-frame for the onset of
42 UHP exhumation across the NW Himalaya. Furthermore, our regional synthesis demonstrates a
43 causative link between changes in the subduction dynamics of the India-Asia collision zone at 47-46
44 Ma and the resulting mid-Eocene plate network reorganization. The onset of UHP exhumation therefore
45 provides a tightly constrained time-stamp significant geodynamic shifts within the orogen and wider
46 plate network.

47

48 **Key words:** U-Pb geochronology; Himalayan Orogeny; eclogite; UHP metamorphism; continental
49 collision; subduction-exhumation

50 **1. Introduction**

51 The exhumation of continental crust from ultra-high pressure (*hereafter referred to as "UHP*
52 *exhumation"*) is an important yet poorly constrained orogenic process (Guillot et al., 2009, Hacker and
53 Gerya, 2013, Warren, 2013). Many studies relate UHP exhumation to other orogenic processes such as
54 a reduction in the slab-pull force during the subduction of continental crust, slab break-off, and/or onset
55 of various modes of orogenic extension/collapse (e.g., Brun and Faccenna, 2008, Yamato et al., 2008,
56 Guillot et al., 2009, Hacker et al., 2010, Little et al., 2011, Burov et al., 2014, Chen et al., 2022). As
57 such, UHP exhumation in the rock record is often interpreted to signify an important shift in the tectonic
58 and geodynamic regime of an orogen (e.g., O'Brien et al. 2001; Yamato et al., 2008, Guillot et al., 2009,
59 Hacker et al., 2010, Soret et al., 2021, Chen et al., 2022), However, the exact relationships between
60 continental UHP exhumation and the geodynamics of orogenesis and plate tectonics, as well as the
61 mechanisms that facilitate this process, remain poorly constrained due to the difficulty of investigating
62 UHP metamorphism and exhumation from the bedrock record (Hacker and Gerya, 2013, Warren, 2013,
63 O'Brien, 2019). To better understand the processes of continental UHP exhumation and its impacts on
64 orogenesis and plate tectonics, we must combine petrology-based pressure-temperature (P-T) pathways
65 with radioisotope geochronological constraints. This approach allows us to determine the timing and
66 rates of UHP exhumation through P-T space and with respect to the orogeny. However, such attempts
67 to constrain the timing and rates of UHP exhumation are always complicated by the uncertainty with
68 which accessory phase ages are related to P-T paths (e.g. Kohn et al., 2017, O'Brien, 2019a). This

69 problem is clearly demonstrated in the UHP Tso Morari Complex of the NW Himalaya (Steck *et al.*,
70 1998; Epard and Steck, 2008; Guillot *et al.*, 2008), which forms the central study region of our
71 investigation.

72 The Tso Morari Complex is one of two UHP terranes in the NW Himalaya, the other being the Kaghan
73 Valley Complex, located ~450 km to the west of Tso Morari (Steck *et al.*, 1998; O'Brien *et al.*, 2001;
74 Parrish *et al.*, 2006; Epard and Steck, 2008; Guillot *et al.*, 2008; Buchs and Epard, 2019). Over the last
75 three decades, geochronological studies of Tso Morari have yielded a range of estimates for the timing
76 of UHP metamorphism and exhumation, spanning a period of 11 Ma (Leech *et al.*, 2007; Donaldson *et*
77 *al.*, 2013; St-Onge *et al.*, 2013). Early interpretations were based on a limited number of spot analyses
78 of zircon (e.g. n=19, Leech *et al.*, 2007), whereas the later publication by Donaldson *et al.*, (2013)
79 provided a more robust record (n = 108) with a zircon age distribution ranging from 53 to 37 Ma, with
80 a peak at 47 – 43 Ma that provided a substantially narrowed estimate of peak UHP metamorphism. In
81 contrast, estimates from the Kaghan UHP eclogites are tightly constrained to ~46 Ma (Kaneko *et al.*,
82 2003; Parrish *et al.*, 2006; Zhang *et al.* 2022). From these constraints, a diverse but poorly constrained
83 set of models have been proposed for the exhumation of Tso Morari and Kaghan Valley complexes
84 (Schwartz *et al.*, 2007; Kylander-Clark *et al.*, 2008; Möller *et al.*, 2015; Boutelier and Cruden, 2018).

85 The spread of ages for UHP metamorphism and exhumation in Tso Morari, could, reflect differences
86 in analytical techniques and their inaccuracies, but a likely cause is also the difficulty and uncertainty
87 associated with linking geochronological data with independently constrained metamorphic petrology
88 (Foster and Parrish, 2006; Kohn *et al.*, 2017, O'Brien, 2019a). This problem is compounded further in
89 the Tso Morari complex because UHP rocks exist as metre-scale eclogite facies mafic pods hosted
90 within amphibolite facies felsic gneiss. Accessory phases suitable for geochronology are rare in eclogite
91 pods, so some studies have attempted to relate the P-T evolution of eclogite pods with geochronological
92 constraints from the amphibolite facies gneiss (St-Onge *et al.*, 2013). However, this approach carries
93 additional uncertainties surrounding the structural and metamorphic relationships between the eclogite
94 pods and the amphibolite facies gneiss in which they are hosted (O'Brien, 2018).

95 To overcome these problems, estimates of the timing of UHP exhumation are better resolved using a
96 combination of modern petrochronology techniques (Kohn *et al.*, 2017), and where mineral size and
97 zonation allows, high precision techniques (Parrish *et al.*, 2006). The former can utilise a range of
98 approaches, including: 1) the combination of geochronology, trace element geochemistry and
99 metamorphic petrology, to quantitatively relate precise ages to specific stages on a metamorphic P-T
100 path (e.g. Rubatto, 2002; Rubatto and Hermann, 2003); and 2) identifying and selecting a range of
101 accessory phases associated with different metamorphic assemblages and lithologies, which
102 collectively span a wide range of closure temperatures (Regis *et al.*, 2016; Lotout *et al.*, 2018; Tual *et*

103 al., 2022). Such techniques provide the best opportunity to accurately and precisely constrain the timing
104 of UHP exhumation.

105 In this study, we employ a range of petrochronological techniques to precisely constrain the timing of
106 UHP metamorphism and exhumation of the Tso Morari Complex. We use detailed petrographic
107 analyses, including major and trace element x-ray maps of garnet, to identify and relate prograde, peak,
108 and retrograde metamorphic assemblages in mafic eclogite samples and amphibolite facies felsic gneiss
109 samples (the Puga gneiss). This allows us to select a variety of accessory phases from different
110 metamorphic assemblages for U-Pb geochronology, to constrain the timing of metamorphism at
111 different points of the P-T path. Using laser ablation inductively coupled plasma mass spectrometry
112 (LA-ICP-MS), we analyse zircon and rutile from an eclogite pod sample, and zircon, rutile, and
113 xenotime from a felsic gneiss sample to constrain the timing of prograde, peak, and retrograde
114 metamorphism. In addition, we analysed zircon from the eclogite using chemical abrasion isotope
115 dilution thermal ion mass spectrometry (CA-ID-TIMS) to more precisely constrain the timing of zircon
116 crystallization. Our results suggest that the onset of exhumation of Indian continental crust from UHP
117 conditions occurred synchronously across the Tso Morari complex and Kaghan Valley complex at ~47-
118 46 Ma.

119 **1.2 The geodynamic significance of UHP exhumation**

120 After presenting our new geochronology data and interpreting its implications for the metamorphic
121 evolution of the Tso Morari complex, we then consider the geodynamic implications of our new
122 constraints for the Himalayan orogeny and the wider tectonic plate network. The early stages of
123 continental collision involve a shift in the geodynamics of subduction due to the positive buoyancy of
124 continental crust on the down-going plate (Afonso and Zlotnik, 2011, Hacker and Gerya, 2013, Warren,
125 2013, Capitanio et al., 2015, Laik et al., 2022). The exhumation of continental crust from UHP
126 conditions provides an important marker of this geodynamic shift, due to the association of UHP
127 exhumation with modification(s) to the subduction dynamics of a collision zone such as slab break-off,
128 a reduction in slab dip, and/or an increase in mechanical coupling between the upper and lower plates
129 (e.g., Guillot et al., 2008, Afonso and Zlotnik, 2011, Warren, 2013, Chen et al., 2022). The timing of
130 UHP exhumation is especially important, as this constraint allows scientists to interpret the structural,
131 metamorphic, and magmatic evolution of orogenesis with respect to changes in the subduction
132 dynamics of a collision zone and vice-versa, offering an invaluable opportunity to test geodynamic and

133 plate tectonic models of continental collision against bedrock constraints (e.g., Guillot et al., 2008,
134 Yamato et al., 2008, Capitanio et al., 2015, Chen et al., 2022).

135 More broadly, the timing of continental UHP exhumation can be used to better understand the impacts
136 of continental collision on tectonic plate networks. Continental collisions change the summative slab
137 pull forces within their encompassing plate network, either by reducing slab pull of the collision zone
138 during subduction of buoyant continental lithosphere or by removing slab pull from the collision zone
139 altogether via slab break off (e.g., Afonso and Zlotnik, 2011, Capitanio et al., 2015, Chen et al., 2022,
140 Laik et al., 2022). In some cases, this can trigger a plate network reorganization event, defined by a
141 short period (~10-20 Myr) of rapid and widespread modifications to the speed and direction of tectonic
142 plate motions and the topology of plate boundaries (e.g., Matthews et al., 2012, Gibbons et al., 2015,
143 Müller et al., 2016, Gürer et al., 2022). Understanding the causes and occurrences of these
144 reorganizations is important as they can drastically alter the course of Earth's paleogeographic and
145 tectonic evolution in a relatively short period of time.

146 Our understanding of plate network reorganizations is limited due to the difficulty of robustly
147 demonstrating a causative relationship between temporally overlapping, but spatially disparate tectonic
148 events and coeval changes in plate kinematics. This is particularly problematic when investigating
149 reorganizations associated with continental collision, due to the complexity of their bedrock record and
150 the uncertainties of geochronological constraints that obfuscate the duration and order of localized
151 events and their geodynamic relationships with network-wide modifications of plate kinematics. Faced
152 with this problem, constraints such as the timing of UHP exhumation, which directly correspond to
153 changes in subduction dynamics of a collision zone, can provide crucial geochronological markers for
154 constraining causative relationships between collisions and plate network reorganizations (e.g.,
155 Matthews et al., 2012, Gibbons et al., 2015).

156 A major plate network reorganization for the Indian, Eurasian, African, Australian, and Antarctic plates
157 is recognized by previous studies during the Eocene (Gibbons et al., 2015). This reorganization is
158 arguably associated with the Himalayan orogeny (Gibbons et al., 2015, Zahirovic et al., 2016), but
159 uncertainty in the collision history of the orogeny (e.g., Hu et al., 2016; van Hinsbergen et al., 2019;
160 Kapp & DeCelles, 2019; Parsons et al., 2020) means that a precise and unambiguous timestamp linking
161 the bedrock record of the orogen with the Eocene plate network reorganization has been lacking. It is
162 therefore unclear exactly how the structural and metamorphic evolution of the Himalayan orogeny
163 relates to and/or caused plate kinematic changes during the Eocene reorganization. Our new, precise
164 geochronometric constraints for the UHP exhumation in the NW Himalaya present a valuable
165 opportunity to evaluate the temporal relationships between the Himalayan orogeny and the Eocene plate
166 network reorganization and consider their causative links. With the modern advent of the user-friendly
167 GPLates plate reconstruction application (Müller et al., 2018), similar integrated analyses of collisional

168 metamorphism and plate kinematics could be used to better understand the plate tectonic consequences
169 of other orogenies.

170 **2. Geological Background**

171 **2.1. The Himalayan orogeny and the India-Asia collision: Definitions**

172 In order to interpret our data with respect to the Himalayan orogeny and India-Asia collision, it is
173 necessary to outline some definitions used hereafter. Working hypotheses for tectonic formation of the
174 Himalayan orogeny and the India-Asia collision can be broadly categorized into “Single Collision” and
175 “Double Collision” models (e.g., Hu et al., 2016; Kapp and DeCelles, 2019, Parsons et al., 2020). Single
176 Collision models propose a single continental collision between India and Eurasia, beginning at ~61
177 Ma (An et al., 2021) with convergence and under thrusting continuing to the present day (e.g., Gansser,
178 1966; Le Fort, 1975; Hu et al., 2016, Ingalls et al., 2016). Such models are not considered tenable as
179 they require extreme volumes of continental subduction and do not adequately explain the significant
180 kinematic and geodynamic changes which occur within the orogen and the surrounding plate network
181 between 50-40 Ma (e.g., van Hinsbergen et al., 2019, Parsons et al., 2020, Parsons et al., 2021). As
182 such, the results of this study are interpreted in the context of the competing “Double Collision” models,
183 for which two alternative working hypotheses exist, differing with respect to the nature of the first
184 collision. In both hypotheses, “first collision” began at ~61 Ma (e.g., Hu et al., 2015; An et al., 2021)
185 but corresponds, to either (1) collision of the Indian continent with an equatorial Neotethys intra-oceanic
186 arc (e.g., Patriat & Achache, 1984; Stampfli and Borel, 2004; Bouilhol et al., 2013; Replumaz et al
187 2014; Burg and Bouilhol, 2019); or (2) collision between an India-derived *microcontinent* and the
188 Eurasian active margin (e.g., Sinha Roy, 1976; van Hinsbergen et al., 2019; Zhou and Su, 2019).
189 “Second collision” occurred between the Indian continent and the Eurasian active margin sometime
190 between 50 Ma to 25 Ma (e.g., Patriat & Achache, 1984; Replumaz et al 2014; Burg and Bouilhol,
191 2019, Searle, 2019, van Hinsbergen et al., 2019, Parsons et al., 2021).

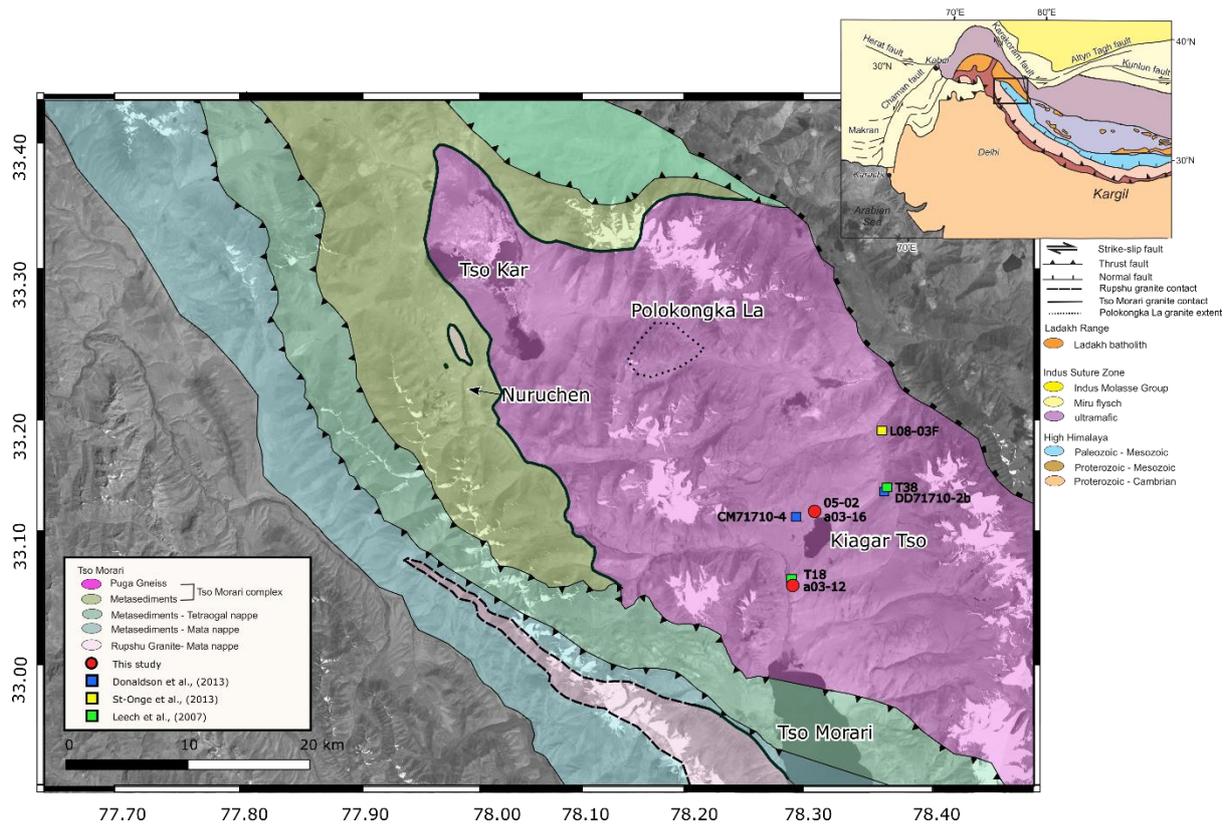
192 In the context of these double collision models, the Himalayan orogeny corresponds to the deformation
193 of Indian continental crust which initiated during first collision and continued through second collision.
194 In contrast, the “India-Asia collision” *sensu-stricto* corresponds to the second collision event only (e.g.,
195 Parsons et al., 2020; 2021). Debate continues to surround the relative validity of the two “double
196 collision” hypotheses and further considerations can be found in recent reviews (e.g., Kapp and
197 DeCelles, 2019, Searle, 2019, van Hinsbergen et al., 2019, Parsons et al., 2020); our study is presented
198 in the context of both double collision models (e.g., Burg & Bouihol 2019 versus van Hinsbergen et al.
199 2019) as it is beyond the scope of the new data presented in this paper to address their relative validity.

200 **2.2. Geology of the Tso Morari Complex**

201 The Tso Morari dome is situated on the north-western margin of the Indian plate (Fig. 1), and is
202 separated from the Ladakh batholith of the Asian plate located to the north by the Indus suture zone
203 (Fuchs and Linner, 1996). The Tso Morari dome comprises a set of stacked nappes, folded into a
204 northwest-southeast trending periclinal antiform. (Steck et al., 1998; Buchs and Epard, 2019). The
205 structurally lowermost nappe is the Tso Morari Complex (also known as the Tso Morari Gneiss, Epard
206 & Steck, 2008), which crops out in the centre of the dome. The Tso Morari Complex contains
207 Ordovician granite and disrupted mafic dykes and sills intruded into Cambrian sediments of the Indian
208 continent. Granitic rocks are variably deformed, resulting in an array of undeformed metagranites,
209 augen gneiss and garnet-mica-schists within the Tso Morari Complex. The range of deformation states
210 preserves different parts of the P-T-t history of the Tso Morari Complex. The early subduction-related
211 history is rarely preserved except as thin corona textures in low strain metagranite (Bidgood et al, 2022),
212 whereas the high strain gneisses are often overprinted by later, amphibolite facies metamorphism and
213 exhumation-related deformation. These end-member states are distinguished locally as the Polokongka
214 La granite and the Puga Gneiss, which share the same granitic protolith, and differ only in their state of
215 strain and metamorphic evolution (Girard and Bussy, 1999).

216 Mafic rocks within the Tso Morari complex locally preserve eclogite-facies mineral assemblages
217 formed at conditions of > 26 kbar, $500-645^{\circ}\text{C}$ (e.g. de Sigoyer and Guillot, 1997; Guillot et al., 1997;
218 St-Onge et al., 2013; Bidgood et al., 2020). Evidence of high-pressure metamorphism is rarely observed
219 in the felsic metagranitoid rocks but has been recorded by glaucophane-bearing metasediments (Guillot
220 et al., 1997), pseudomorphs after coesite within the Polokongka La granite (Bidgood et al., 2020) and
221 thin corona textures (garnet, phengite, ilmenite) and pseudomorphs (kyanite, garnet and phengite) after
222 cordierite in low strain metagranites (Bidgood et al, 2022). Later overprinting at amphibolite-facies
223 conditions is recorded in the retrogressed eclogites at $610 \pm 30^{\circ}\text{C}$ at 9 ± 3 kbar (de Sigoyer et al.,
224 1997) and in the felsic rocks at $650 \pm 50^{\circ}\text{C}$ at 9 ± 1 kbar (Girard, 2001), $630^{\circ}\text{C} \pm 30^{\circ}\text{C}$ at 9 ± 2 kbar
225 (Guillot et al., 1997) and $725 \pm 50^{\circ}\text{C}$ at 7.1 ± 1.0 kbar (St-Onge et al., 2013).

226 This study describes the record of subduction-related major and accessory mineral growth preserved in
227 a rare garnet-bearing Puga Gneiss which records a part of the P-T-t history not preserved elsewhere.
228 Additionally, new data is presented which tightly constrains zircon growth in the mafic eclogites, which,
229 in conjunction with the observations in the Puga Gneiss, allows us to place the petrographic
230 observations in a geochronologic and geodynamic context.



231

232 *Figure 1.* Geological map adapted from Epard and Steck (2008), St-Onge et al. (2013) and references
 233 therein, showing location of geochronological samples. Background USGS landsat data downloaded
 234 from <https://earthexplorer.usgs.gov/>. Colouring corresponds to bands 762, greyscale by luminosity.

235 2.3 Geochronology of the Tso Morari Complex

236 The range of estimates for UHP metamorphism and exhumation in the Tso Morari complex (54 – 44.5
 237 Ma, Leech et al., 2007; St-Onge et al., 2013; Donaldson et al., 2013) is less tightly resolved than the
 238 age estimates for Kaghan (~46 Ma, Kaneko et al., 2003; Parrish et al., 2006; Zhang et al., 2022). Given
 239 the differences in analytical techniques used to date the Tso Morari complex and the difficulty in
 240 relating these dates to metamorphic stages, it remains unclear to what degree the spread of ages from
 241 Tso Morari can be assigned to geological heterogeneity or to analytical uncertainty and artefacts. For
 242 example, the absence of petrographic context in some studies makes it difficult to interpret the age
 243 clusters with respect to specific metamorphic conditions (see O'Brien, 2006 for further discussion).
 244 Additionally, the link between zircon growth and fluid infiltration (e.g. Chen et al., 2010; Kohn et al.,
 245 2015; Chang & Zhang, 2017; Shezovatov et al., 2021) has not been explored in the context of the
 246 published data. Importantly, given the types of data yielded from Tso Morari thus far (see following
 247 summary), it is not possible to robustly constrain whether zircon growth in the mafic eclogites was
 248 prolonged or punctuated, and whether the range of dates correspond to prograde, peak and/or retrograde
 249 metamorphic conditions. Given the narrow age range for UHP metamorphism from Kaghan only 450
 250 km away, comparative age estimates from Tso Morari would have significant implications for the

251 processes of continental subduction and UHP exhumation; as such, this is the primary motivation for
252 this study. Here we present new and precise ages of eclogite-facies metamorphism that can be compared
253 with the ages recorded at Kaghan, along with new in situ data which, when combined with the
254 petrographic interpretation, records an additional petrographic age and stage of the evolution of the Tso
255 Morari complex that has not been previously observed.

256 The Puga Gneiss and Polokongka La granite contain zircon with thin metamorphic rims surrounding
257 igneous cores, with the latter dated by U-Pb at 479 ± 1 Ma (Girard and Bussy, 1999; Leech et al., 2007).
258 Initial estimates of the age of high pressure metamorphism by De Sigoyer et al. (2000) used Lu-Hf
259 isochron ages on garnet-clinopyroxene-whole-rock data, Sm-Nd on garnet-glaucophane-whole-rock
260 data, and U-Pb in allanite, giving ages of 45 ± 4.4 Ma, 55 ± 12.0 Ma and 55 ± 17.0 Ma, respectively.
261 These estimates attempted to date minerals which can definitively be linked to metamorphic
262 reactions/conditions, but this was hindered by the large uncertainties on these dates.

263 A spread in zircon dates from 55 to 45 Ma was recorded in the Puga Gneiss by Leech et al (2007) by
264 U-Pb SHRIMP dating of the metamorphic rims found around igneous zircon grains ($n = 19$; 4 ages
265 interpreted). Sub-concordant data were split into groups arbitrarily (see O'Brien, 2006), and interpreted
266 to represent ultrahigh-pressure metamorphism at 53.3 ± 0.7 Ma, followed by further zircon growth in
267 eclogite facies conditions at 50.1 ± 0.6 Ma, amphibolite facies at 47.5 ± 0.6 Ma, and a lower temperature
268 growth at 45.2 ± 0.7 Ma. St-Onge et al. (2013) recorded monazite and allanite U-Pb SHRIMP ages of
269 45.3 ± 1.1 Ma and 43.3 ± 1.1 Ma respectively, interpreted to represent post-eclogite facies peak
270 temperature metamorphism of 7.0 – 8.4 kbar at 705 – 755 °C, based on pseudosection modelling of the
271 observed garnet breakdown reaction. Cooling through phengite, biotite and muscovite Ar-Ar closure
272 temperatures is recorded at 48.0 ± 2.0 Ma, 31.1 ± 0.3 and 29.3 ± 0.3 Ma (De Sigoyer et al., 2000)
273 respectively, with further cooling recorded by apatite and zircon fission track data at $\sim 23.5 - 7.5$ Ma
274 (Schlup et al., 2003). Dates from within the Puga Gneiss are dominated by post-peak, amphibolite facies
275 ages recording Barrovian metamorphism, cooling and uplift.

276 The mafic eclogites better preserve the early high-pressure metamorphic history in their major and
277 accessory mineral assemblages. St-Onge et al (2013) analysed zircon in situ with SHRIMP U-Pb
278 geochronology, yielding an age of 58.0 ± 2.2 Ma ($n = 2$) for zircon included in the core of a garnet, and
279 50.8 ± 1 Ma ($n = 4$) for zircon included/adjacent to matrix barroisite, phengite and garnet. The older
280 age is interpreted to record zircon crystallisation during prograde garnet growth to high pressure, and
281 the younger age is interpreted to represent peak metamorphism in the eclogite facies. Whilst the St.
282 Onge et al. (2013) study provides a textural constraint on the prograde zircon ages which was previously
283 unknown, the data include a limited number of spots and the presence of common lead. Donaldson et
284 al. (2013) used in situ split-stream LA-ICP-MS to measure U-(Th)-Pb combined with REE
285 concentrations of zircon from two mafic eclogites, located ~ 10 km apart. Lower intercept dates of each

286 sample overlap at 45.3 ± 1.6 Ma and 44.2 ± 1.2 Ma. These ages broadly overlap with that of Kaghan,
287 as highlighted by Donaldson et al. (2013), but these authors also argue for protracted zircon
288 crystallisation in the eclogite facies (ca. 47 – 43 Ma) based on the spread of single spot dates and
289 consistent REE signatures with an absence of an Eu anomaly and flat HREE profiles. The Donaldson
290 et al. (2013) results improve upon the constraints of St. Onge et al. (2013) based on the number of data,
291 but are still hampered by both the presence of common lead and the precision of the in situ method
292 which yields single spot uncertainties of ca. 1 to 10 Ma.

293 The age of UHP metamorphism in the Kaghan Valley Complex has been estimated using U-Pb
294 SHRIMP and U-Pb ID-TIMS analyses of zircon from eclogite-facies mafic rocks, yielding ages of 46.2
295 ± 0.7 Ma (Kaneko et al., 2003) and 46.4 ± 0.1 Ma (Parrish et al., 2006), respectively. These zircons
296 were found included in UHP garnet rims with coesite inclusions. Eclogite facies ages of zircon and
297 allanite from Kaghan were also estimated using U-Pb and Th-Pb ID-TIMS analyses at 45.5 ± 6.6 Ma
298 and 46.5 ± 1.0 Ma, respectively (Parrish et al., 2006). An additional age of eclogite facies zircon was
299 estimated using U-Pb SIMS at 46 ± 2 Ma in Naran, 30 km south-west of Kaghan (Zhang et al., 2022).
300 A compilation of geochronology of high pressure metamorphism in the north west Himalaya can be
301 found in Supporting Information 1 and is compared to our new data in Figure 8.

302 **3. Petrography**

303 **3.1. Analytical methods**

304 Petrographic study of 29 mafic eclogites and 28 Puga Gneiss samples from the Tso Morari Complex
305 was undertaken, with one fresh eclogite (a03-16), one retrogressed eclogite (a03-12) and one gneiss
306 sample (05-02) selected for further petrological analyses. Major element compositions of minerals that
307 exhibit solid solutions were measured using a Cameca SX-5 field emission electron microprobe at the
308 University of Oxford, with a 15 kV acceleration potential, 20 nA beam current, 30 second count time
309 per major element (30 second background) and 60 second count time on Ti (60 second background). A
310 range of natural and synthetic oxide standards were used including albite (Na, Al, Si), Orthoclase (K),
311 MgO (Mg), wollastonite (Ca), andradite (Fe), Mn metal (Mn) and synthetic TiO₂ (Ti) and analyses
312 were verified against secondary mineral standards. Mineral spot analyses and line profiles were taken
313 across garnet to determine the extent of intracrystalline compositional variation (see Supporting
314 Information 2).

315 Quantitative major element X-ray maps of garnets were collected from polished thin sections using the
316 CAMECA SX-5 field emission electron microprobe at the University of Oxford at a working distance
317 of 10 mm, a 15 kV acceleration potential, 170 nA current, 0.06 s dwell time and a 3 μ m step size for
318 the elements P, Ca, Mn, Mg and Fe. A dwell time of 0.032 s and a 200 nA beam current were used for

319 the elements Al, Si, Ti, Y and Yb. A range of natural and synthetic oxide standards were used including
320 Durango apatite (P), andradite (Ca, Fe), Mn metal (Mn), MgO (Mg), albite (Al, Si), TiO₂ (Ti), Y metal
321 (Y), Yb metal (Yb).

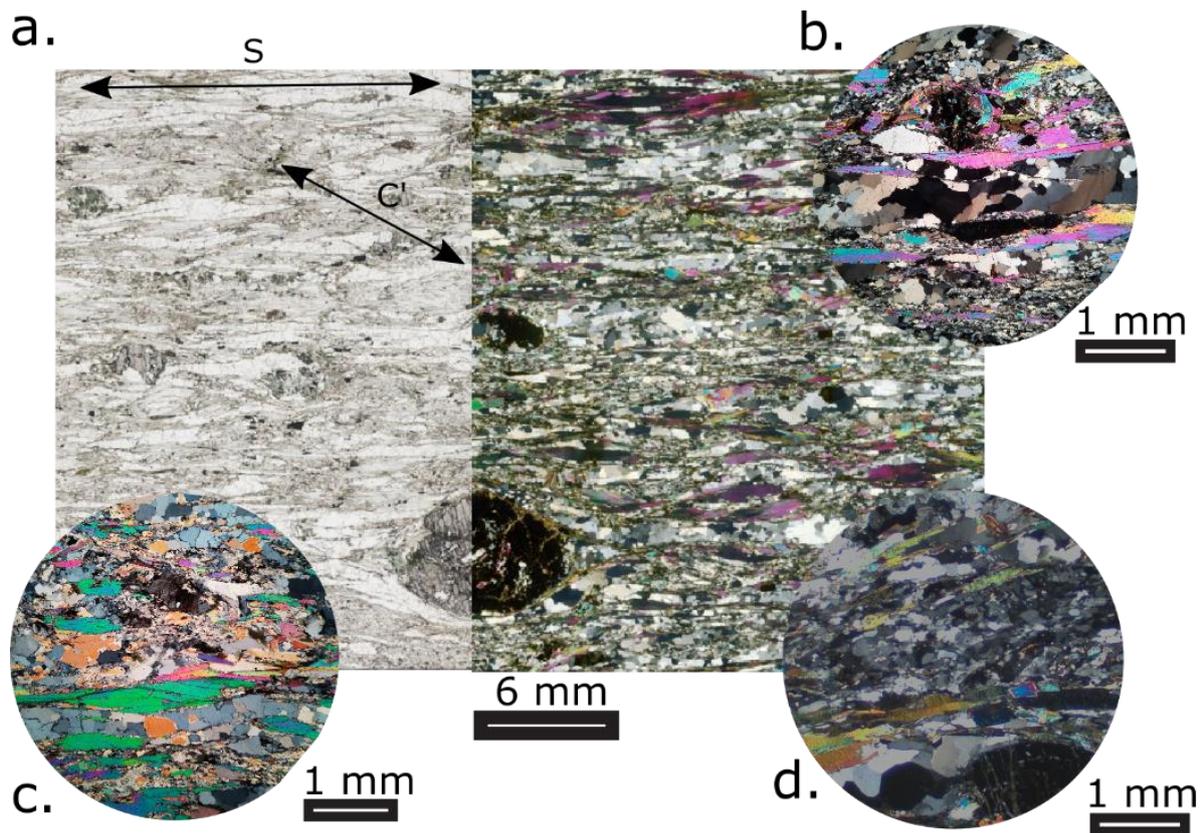
322 **3.2. Petrography and petrology: Observations**

323 **3.2.1. Puga Gneiss (sample 05-02)**

324 Puga Gneiss sample 05-02 was collected from the north shore of Kiagar Tso (33.1214°N, 78.2958°E),
325 in the middle of the Tso Morari Complex. Sample 05-02 is a strongly-foliated, garnet-bearing gneiss
326 comprising albite, quartz, muscovite, biotite and garnet, with accessory zircon, apatite, rutile, and
327 xenotime. Sample 05-02 shows a schistosity (S₂) defined by bands of white mica, albite and quartz
328 (Fig. 2a) which wrap around larger garnet porphyroblasts. Quartz occurs in polycrystalline ribbons or
329 lenses separated by narrow bands of white mica, indicative of a high-strain fabric (Fig. 2a-d). The quartz
330 ribbons are cut by discontinuous shear bands forming an S-C' fabric. Quartz grains are > 200 μm and
331 equant with amoeboid shapes along grain boundaries indicative of grain boundary migration (GBM)
332 dynamic recrystallisation (e.g. Stipp et al., 2002). Albite bands and lenses are largely composed of fine-
333 grained aggregates. Among these, larger feldspar grains show subgrains of similar size to the dominant
334 population, suggesting that dynamic recrystallization occurred in the subgrain rotation (SGR) regime
335 for feldspar (Passchier & Trouw, 2005). Zircon and rutile occur in the matrix.

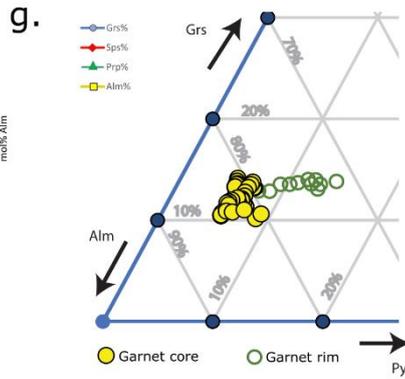
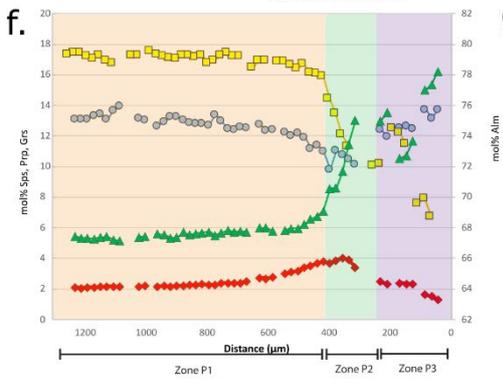
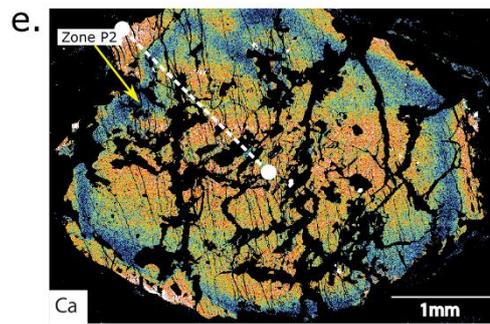
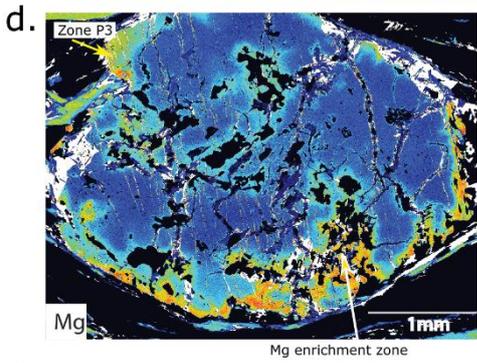
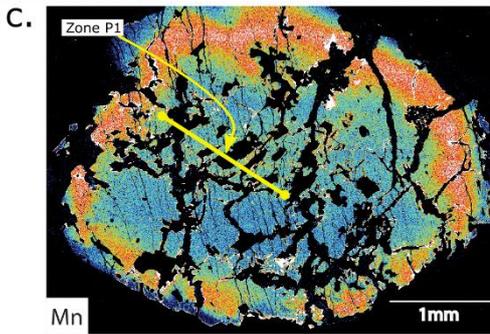
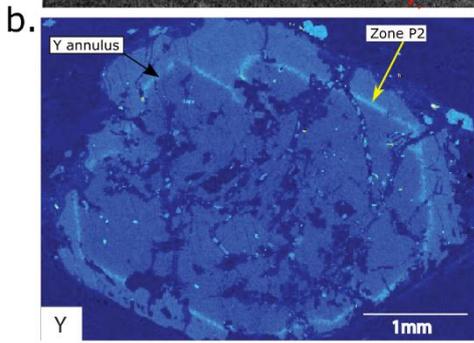
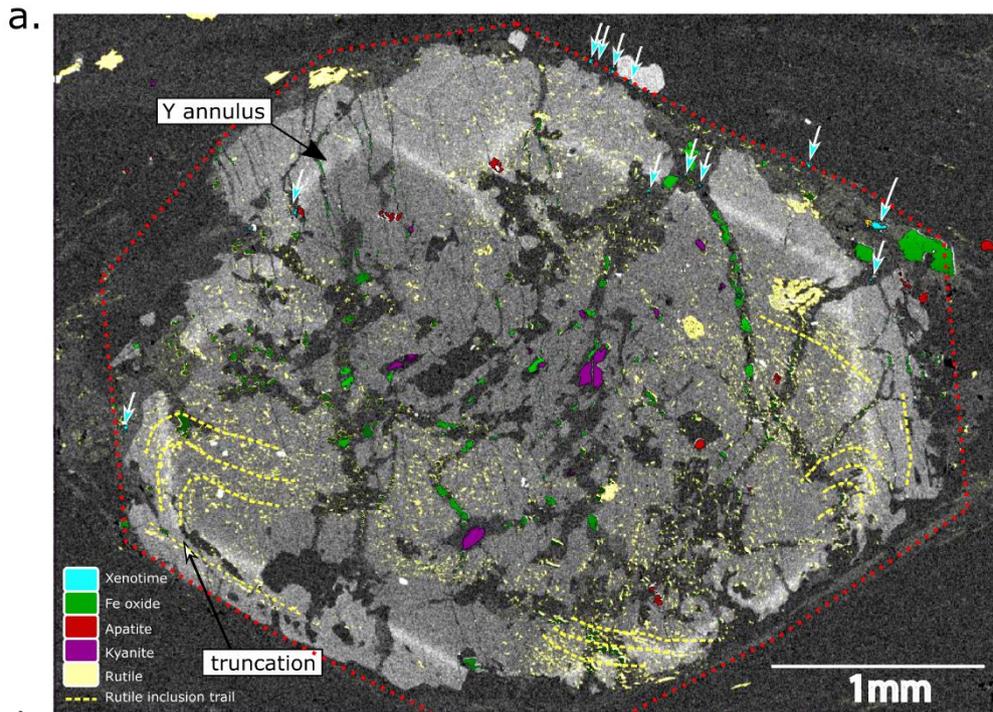
336

337 Garnet porphyroblasts of varying size are scattered through the rock. They are abundantly fractured,
338 and partly replaced by biotite, white mica and chlorite. The studied section contains one large (6 x 5
339 mm) ovoid garnet porphyroblast (Fig. 3) that contains significant textural complexity (see below). Fine-
340 grained inclusion trails of quartz and rutile define a primary foliation (S₁), which is oblique to the
341 matrix fabric (S₂) and folded at the core-rim boundary (Fig. 3a). Kyanite occurs within the garnet core,
342 adjacent to staurolite, surrounded by white mica (Fig. 3a). The garnet is subhedral with embayments
343 containing biotite, white mica and xenotime (Fig. 3a).



344

345 Figure 2. a. Puga Gneiss Sample 05-02. PPL and XPL image showing gneissose texture consisting of
 346 quartz-rich, feldspar-rich and mica-rich layers. Large garnet wrapped by fabric (S2) shown in Figure 3.
 347 b-c Puga Gneiss Sample 05-01, adjacent to 05-02. b. Polycrystalline quartz-rich band in between mica
 348 sheets. Recrystallised quartz shows undulose extinction. Quartz grains in the lower part of the image
 349 span the width of the mica ribbons. c. Stacked white mica sheets curve into high strain fabric showing
 350 top to the right sense of shear. Annealed quartz band and dusty albite and apatite present. d. Puga Gneiss
 351 Sample 05-02. Fine-grained albite lens between mica sheets.



353 Figure 3. Puga Gneiss Sample 05-02 garnet. a. The brightest grains from the Ti map (yellow), Ca map
354 (red), Fe map (green), Y map (cyan) and Al map (purple) are overlain on an Y map adjusted for Y-
355 zonation in garnet. The location of xenotime grains can be identified based on the brightest Y locations
356 to correspond with the micaceous rim after garnet. The map also highlights the orientation of rutile
357 inclusion trails and the location of rectangular rutile clusters. The approximate outline of the original
358 garnet is highlighted as a red dashed line. b-e. Yttrium, manganese, magnesium and calcium EPMA
359 element maps. Location of the core-rim boundary defined by the yttrium high. f. Garnet zoning profiles
360 from core (left) to rim (right). g. Ternary diagram of transect across garnets. Approximate location of
361 garnet transect shown by white dashed line on panel e.

362

363 Garnet exhibits concentric zonation, comprising three distinct zones (Fig. 3). The garnet core (zone P1,
364 Fig. 3c) has an approximately uniform composition of $\text{Alm}_{79,0}\text{Grs}_{12,5}\text{Pyr}_{5,5}\text{Spss}_{4,0-2,3}$. Inclusions of
365 quartz, rutile and white mica are found throughout zone 1, whereas inclusions of kyanite are restricted
366 to the inner portion of zone P1 (Fig. 3a,f). Zone P2 (Fig. 3b,d,f) is marked by a decrease in Ca and
367 increase in Mn. It contains a narrow annulus of high Y content, and its outer boundary is faceted. The
368 narrow rim zone P3 (Fig. 3d,f) shows a marked increase in Mg (to 16% pyrope) and Ca (to 13%
369 grossular), with lower Mn. It is discontinuous around the garnet margin. Zones P2 and P3 contain
370 quartz, white mica, and rutile inclusions, but to a lesser extent than in zone 1. Kyanite is absent from
371 zones P2 and P3. In zones P1 and P2, magnesium-enriched haloes are developed along the internal
372 fracture network, surrounding many larger inclusions of white mica and quartz, and connecting with
373 the outer zone of the garnet.

374

375 **3.2.2. Mafic eclogite (sample a03-16)**

376

377 Sample a03-16 is a mafic eclogite taken from the same locality as Puga Gneiss sample 05-02
378 (33.1214°N, 78.2958°E), adjacent to sample CM71710-4 of Donaldson et al. (2013), and displays
379 similar features to mafic eclogites described from other localities in the Tso Moriri Complex (e.g.
380 Jonnalagadda et al., 2017; O'Brien & Sachan, 2000; Palin et al., 2014; St-Onge et al., 2013; Wilke et
381 al., 2015). Sample a03-16 has a medium- to coarse-grained granoblastic texture with a major mineral
382 assemblage of garnet, omphacite, phengite, quartz, and talc, with minor amounts of clinozoisite,
383 amphibole, carbonate, rutile and zircon.

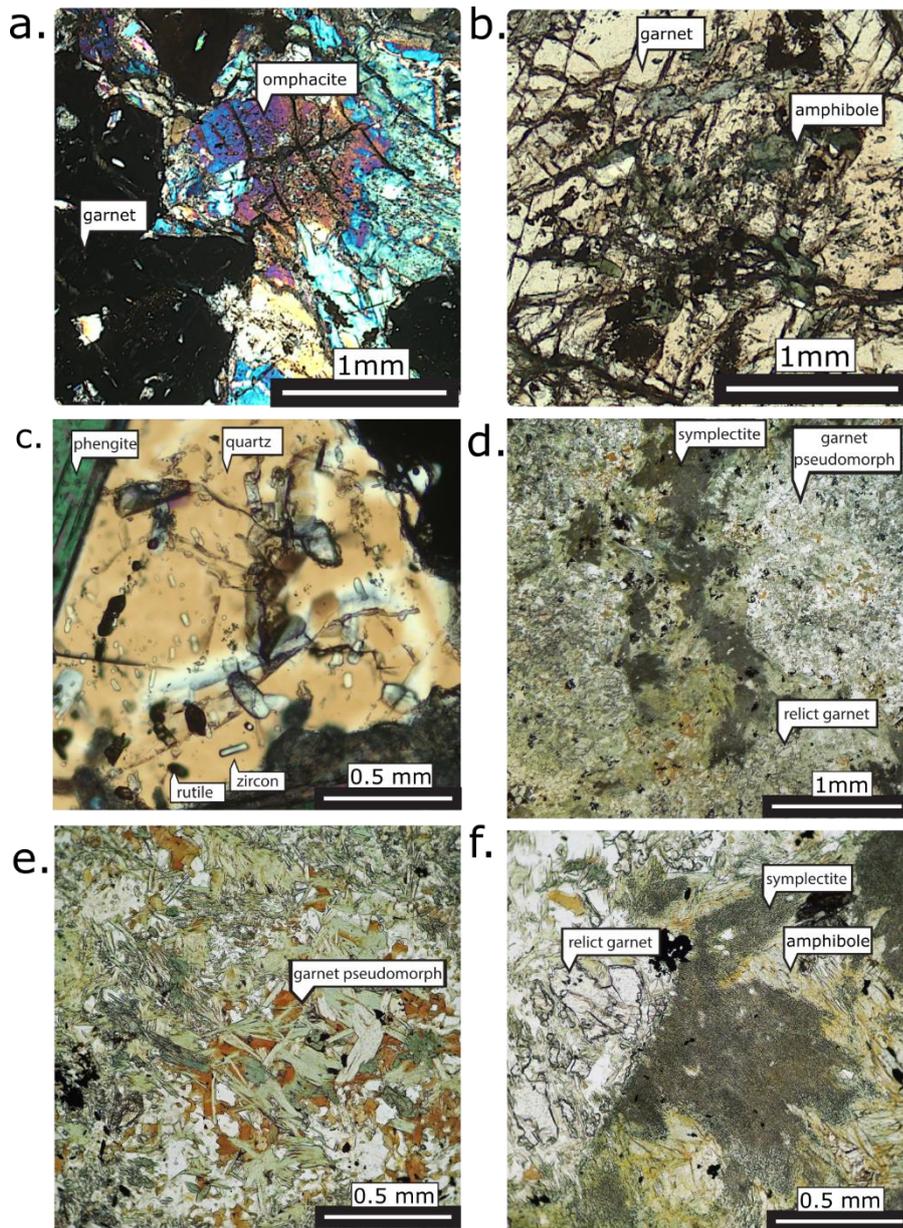
384

385 Garnet and omphacite (Fig. 4a) are in textural equilibrium, forming planar contacts, with coarse grained
386 homogeneous phengite and talc. In some places, symplectites after omphacite are observed, comprised

387 of amphibole, plagioclase and occasional diopside. Dolomite is also present in the matrix as large
388 poikiloblasts containing inclusions of phengite, omphacite and rutile.

389

390 Amphibole occurs in the matrix and as inclusions in garnet (Fig. 4b). In common with other sampled
391 outcrops, coarse-grained matrix amphibole typically has pale green winchite cores, zoned towards a
392 pargasitic composition at rims. Thin mantles of green pargasitic amphibole surround garnet, and
393 amphibole inclusions in garnet are darker blue-green pargasite and ferropargasite. Fine-grained
394 intergrowths of biotite and plagioclase surround phengite. Matrix quartz and omphacite contain clusters
395 of zircon inclusions, whereas garnet, omphacite, quartz, phengite, and talc contain rutile inclusions (Fig.
396 4c). Garnet occurs as subhedral porphyroblasts that are 3-4 mm in diameter (Fig. 4b) and contain
397 inclusions of zoned blue-green amphibole, and also quartz and rutile. Rutile commonly forms angular
398 clusters of grains, both in the matrix and within garnets, which we interpret as replacing a former
399 igneous Fe-Ti oxide.

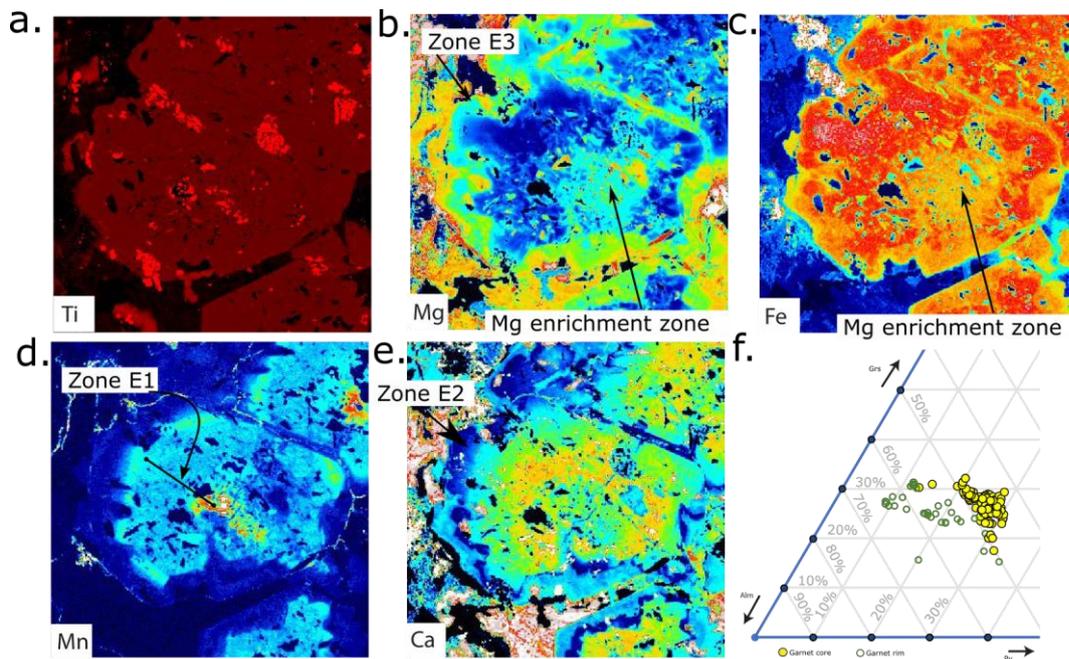


400

401 Figure 4. Photomicrographs of mafic eclogites. a. Sample a03-16, XPL, poikiloblastic garnets and
 402 omphacite. b. Sample a03-16, colour zoning in garnet with amphibole inclusions ranging in colour from
 403 blue to green. c. Sample a03-16, matrix quartz with abundant high relief zircon and dark rutile
 404 inclusions. d-f. Sample a03-12 retrogressed mafic eclogite. d. Subhedral pseudomorphed garnets
 405 present on the edge of the image. Dark-coloured, fine-grained intergrowth of plagioclase and amphibole
 406 (symplectite) after omphacite. Coarse-grained, green matrix amphiboles interlocked with omphacite
 407 pseudomorphs. e. Core of large pseudomorphed garnet showing aggregate of chlorite, biotite and minor
 408 green amphibole set in plagioclase. f. Dark amphibole-plagioclase symplectite after omphacite adjacent
 409 to partly pseudomorphed garnet.

410

411 Major and trace element maps of a single garnet in sample a03-16 show three zones (Fig. 5). Zone E1
 412 (Fig. 5d) defines the core region, which has a composition of Alm48-42 Grs24-29 Pyr20-30 Sps6-1,
 413 with increasing almandine and pyrope components from core to rim (Fig. 5 b and c). The rim of the
 414 zone E1 core is outlined by a Mn-rich annulus. Zone E2 surrounds the zone E1 core and is defined
 415 by a calcium trough with a faceted outline (Fig. 5 d and e). Zone E3 (Fig. 5b) forms along the garnet
 416 rim defined by a Mg-rich, Mn-poor rim with shows an increase in pyrope (to 30-35%) and grossular
 417 (to 30%, decreasing to 16% at the outer rim). Mg-rich E3-like domains also extend in some places
 418 from the outer rim towards the core of the garnet, cutting across concentric garnet zones E1 and E2, and
 419 surrounding inclusions, often forming channelized features (Fig. 5 b and c).



420
 421 Figure 5. Garnet maps from mafic eclogite sample a03-16. a. Ti map showing clusters of rutile
 422 inclusions forming distinct shapes after Fe-Ti oxides. b-e. Mg, Fe, Mn and Ca maps showing concentric
 423 zoning patterns cut by Mg-Fe embayments emanating from the outside rim of the garnet. f. Ternary
 424 diagram of transect across garnet showing Fe-Mg alteration and garnet rim.

425
 426 **3.2.3. Retrogressed mafic eclogite (sample a03-12)**

427
 428 Sample a03-12, along with the adjacent sample a03-09, is a retrogressed mafic eclogite from ‘The
 429 Bridge’ locality (33.0677°N, 78.2758°E), ~8 km south of sample a03-16. This is the same locality
 430 sampled and analysed by Leech et al. (2007) (their sample T18). The sample is composed of garnet,
 431 amphibole, plagioclase, biotite, chlorite, ilmenite, rutile and zircon. Relict garnet is preserved within
 432 pseudomorphs that preserve subhedral garnet shapes, but are largely replaced by plagioclase,
 433 hornblende, biotite and chlorite aggregates (Fig. 4 d and e). Former matrix omphacite has been

434 overprinted by coarse-grained blue-green sodic-calcic amphibole (cf. Palin et al., 2014), and
435 subsequently pseudomorphed by fine-grained amphibole-plagioclase intergrowths (Fig. 4f).
436 Chlorite and biotite form aggregates which overprint the surrounding metamorphic patterns. Rutile
437 grains occur in garnet and matrix amphibole, and are rimmed or replaced by ilmenite in retrogressed
438 areas. Zircon occurs as small grains in the matrix.

439

440 **3.3. Petrography and petrology: Interpretation and metamorphic correlation**

441 **3.3.1. Puga Gneiss (sample 05-02)**

442 The Puga Gneiss (sample 05-02) displays evidence of two metamorphic assemblages. The first is
443 defined by the garnet compositional zoning, and the inclusion suite of quartz, kyanite, rutile and zircon.
444 Quartz and rutile inclusions define a crenulated primary foliation (S1). Kyanite is found exclusively
445 within garnet. The garnet compositional zoning has features that match the zoning in eclogite garnet
446 (cf. St-Onge et al, 2013, and Fig. 5), i.e. a Ca-rich core (P1 = E1), Ca-poor trough (P2 = E2) and Mg-
447 and Ca-rich rim (P3 = E3). This is consistent with prograde growth culminating in eclogite-facies
448 conditions (e.g. St-Onge et al., 2013; Wilke et al., 2015; O'Brien, 2019b). The Mg distribution in the
449 garnet interior implies that some fracturing occurred at near-peak conditions.

450

451 The second assemblage is defined by the rock matrix grains outside the garnet porphyroblasts, which
452 form a segregated quartz-feldspar-mica mineral banding modified by an S-C' fabric. Accessory
453 minerals in this matrix assemblage include rutile and zircon. Dynamic recrystallization of quartz by
454 grain boundary migration suggests a deformation temperature in excess of 530°C (Stipp et al., 2002),
455 and evidence for subgrain rotation in albite is consistent with about 600°C (Passchier & Trouw, 2005).
456 No lower-temperature dynamic recrystallization microstructures are observed, and minor chlorite,
457 generally associated with garnet, is undeformed. Rutile is stable in the rock matrix, commonly enclosed
458 in white mica. Based on these observations, we interpret the matrix assemblage to reflect metamorphism
459 and deformation on the retrograde path, at amphibolite-facies conditions.

460

461 A distinct mineral association surrounds the large garnet in sample 05-02, where parts of the Mg-rich
462 rim zone (P3) of the garnet are missing, due to partial resorption of garnet, and micaceous aggregates
463 in embayments and fractures host xenotime grains (Figure 3a).

464

465 **3.3.2. Mafic eclogite (sample a03-16)**

466 Sample a03-16 contains three identifiable metamorphic assemblages. Garnet and omphacite (Fig. 4a)
467 in textural equilibrium with coarse grained homogeneous phengite and talc define a peak pressure

468 eclogite-facies mineral assemblage (e.g., M2 of St-Onge et al., 2013). Accessory phases within this
469 peak assemblage include zircon, present as clusters of grains included within matrix quartz and
470 omphacite, and rutile, present as inclusions, commonly clustered, within garnet, omphacite, quartz,
471 phengite, and talc (Fig. 4c). Inclusions of zircon and rutile in garnet cores suggest that initial growth
472 of these grains may have begun prior to eclogite facies metamorphism. Within garnet, the increase in
473 pyrope component from core to rim, corresponds to garnet growth during prograde to peak eclogite
474 conditions. We interpret zones E1 to E3 as a prograde to peak assemblage, where zone E3 correlates
475 with the omphacite, phengite and talc peak assemblage described above. The second metamorphic
476 assemblage is defined by the early breakdown products of the peak assemblage phases. The dominant
477 example is the zoned blue-green matrix amphibole, which may represent the product of talc dehydration
478 as well as the influx of external fluids under eclogite-facies conditions (cf. Palin et al., 2014). The third
479 assemblage includes feldspar-bearing symplectites after omphacite, and secondary fine-grained white
480 mica aggregates forming in the irregular rims of coarse-grained phengite. These represent an
481 amphibolite-facies overprint consistent with the matrix assemblage of Puga gneiss sample 05-02.

482

483 **3.3.3. Retrogressed mafic eclogite (sample a03-12)**

484 Sample a03-12 is dominated by a post-peak metamorphic assemblage, whereas the peak assemblage
485 displayed by sample a03-16, is identifiable in a03-12 as relict garnet grains and pseudomorphs after
486 garnet and omphacite. Blue-green amphiboles are interpreted as an early post-peak phase, which grew
487 within the eclogite facies field prior to the appearance of stable sodic feldspar during decompression
488 (e.g., Palin et al., 2014). The rest of the post-peak assemblage is typical of an amphibolite facies
489 retrograde assemblage characterised by fine-grained amphibole and plagioclase intergrowths.
490 Aggregates of chlorite and biotite characterise a lower amphibolite facies overprint. Rutile is found
491 within prograde relict garnets as well as the matrix and has no indication of internal zonation. We
492 therefore interpret rutile as a relict grains from a prograde or peak assemblage, rather than a new phase
493 that nucleated during latest-stage lower amphibolite facies metamorphism. Zircon is present throughout
494 this sample, predominantly as inclusions in peak omphacite and quartz, suggesting that it crystallised
495 at eclogite-facies conditions.

496

497 **3.3.4. Correlation of metamorphic assemblages (M1, M2, M3) in the Puga Gneiss and mafic eclogites**

498

499 Comparison and correlation of the petrography and petrology of samples 05-02, a03-16, and a03-12,
500 allows for the definition of three distinct metamorphic assemblages that reflect distinct portions of the
501 same P-T path of the Tso Moriri complex, experienced and recorded by both lithologies. These
502 assemblages are correlated with the stages defined by St-Onge et al (2013) (M1, M2, M3). Crucially,
503 these assemblages provide a robust means for linking accessory phase geochronology to metamorphic

504 evolution of the Tso Morari complex. These metamorphic assemblages and their constituent accessory
505 phases are summarised as follows:

506

507 *3.3.4.1. M1: Prograde-to-peak eclogite facies assemblage*

508 Zones P1 and P2 of garnet within the Puga gneiss sample 05-02 and zone E1 and E2 of garnet in the
509 mafic eclogite a03-16 define the M1 prograde-to-peak, eclogite facies assemblage. This includes
510 inclusions of kyanite, quartz, white mica, zircon and rutile, within garnet in Puga Gneiss sample 05-02.

511

512 *3.3.4.2. M2: Peak eclogite facies assemblage*

513 High-Mg rims of garnet (zones P3 and E3), plus omphacite, phengite, quartz and talc in mafic eclogite
514 sample a03-16 define the M2 peak eclogite facies assemblage. M2 also includes the high-Mg garnet
515 rims in Puga gneiss sample 05-02. Elsewhere in Tso Morari, similar garnet rim compositions in mafic
516 eclogites contain inclusions of coesite or polycrystalline inclusions after coesite (Sachan et al., 2004;
517 Bidgood et al., 2021). Remnants of this assemblage are also preserved by relict garnet in the retrogressed
518 mafic eclogite sample a03-12.

519

520 *3.3.4.3. M3a/M3b/M3c: Post-peak assemblage*

521 The M3 post-peak assemblage reflects continuing metamorphism from eclogite to lower amphibolite
522 facies conditions, and is subdivided to reflect this. In the mafic eclogite samples (a03-16, a03-12), M3a
523 is recorded by growth of coarse-grained blue-green amphiboles in the eclogite facies, which elsewhere
524 in the Tso Morari Complex, has been linked to the breakdown of talc and the influx of fluid at 23 kbar
525 and 19 kbar respectively (Palin et al., 2014). Post-peak assemblage M3b corresponds to upper
526 amphibolite facies retrograde metamorphism. In the mafic eclogite samples (a03-16, a03-12) M3b is
527 defined by symplectites of fine-grained amphibole and plagioclase intergrowths after omphacite. The
528 lower-temperature association of chlorite with biotite, largely as a replacement of garnet cores in a03-
529 12, is assigned to M3c.

530

531 In Puga gneiss sample 05-02, initial garnet breakdown and the associated nucleation of xenotime,
532 occurred at eclogite to upper amphibolite facies, and therefore correlates with either the M3a or M3b
533 post-peak assemblages observed in the mafic eclogites. M3b is defined by the matrix assemblage of
534 quartz + albite + muscovite, which also contains zircon and rutile, and displays quartz microstructures
535 indicating post-peak deformation temperatures of >530 °C. M3c corresponds to lower amphibolite
536 facies retrograde metamorphism and is represented by overprinting aggregates of chlorite and biotite
537 after garnet.

538

539 **4. U-Pb Geochronology**

540 **4.1. Analytical methods**

541 Zircon grains from the heavy, non-magnetic fraction of sample a03-12 were imaged via
542 cathodoluminescence (CL) using an FEI Quanta 650 environmental scanning electron microscope (E-
543 SEM) at the University of Oxford, using a 10 kV electron beam, 16mm working distance and a beam
544 current of 0.49nA. Zircon grains from sample 05-02 were also mounted on sticky tape in order to
545 analyse the <10 µm thick rims. Rutile grains chosen for analysis were picked from the non-magnetic
546 fraction and were imaged via backscatter electron imaging (using the same E-SEM) to determine the
547 homogeneity of the grains chosen for analysis. None of the rutile grains showed any evidence of zoning.
548 Xenotime was measured in a polished thin section in order to preserve the petrographic relationships
549 observed.

550

551 All geochronology and mineral separation were conducted at the Geochronology and Tracers Facility,
552 British Geological Survey, Nottingham, UK. Laser ablation inductively-coupled plasma mass
553 spectrometry (LA-ICP-MS) was conducted using a Nu Instruments AttoM sector-field single-collector
554 ICP-MS, coupled to an Elemental Scientific Lasers 193nm UC Excimer laser ablation system fitted
555 with a TV2 cell. The method follows that described in Roberts et al. (2016), with uncertainty
556 propagation following recommendations of Horstwood et al. (2016), and age calculation and plotting
557 using IsoplotR (Vermeesch, 2018). Common lead corrected ages, where quoted, use a ²⁰⁷Pb-based
558 method (Chew et al., 2014) and assume a Stacey and Kramers (1975) initial lead composition, and
559 concordance of the final age. All uncertainties are quoted and plotted at 2σ. LA-ICP-MS dates are
560 quoted with propagation of systematic uncertainties, including reference material and decay constant
561 uncertainties; age uncertainties are also expanded to account for the MSWD where the data exhibit
562 overdispersion (Vermeesch, 2018). Accuracy based on zircon and rutile validation materials was better
563 than 1% for zircon and xenotime and 5% for rutile (Plešovice = 339.1 ± 1.9 Ma; GJ-1 = 596.8 ± 2.9
564 Ma; Manangotry = 555.48 ± 4.23; R19 = 465.9 ± 3.9 Ma; Sugluk = 1714 ± 30 Ma; and PCA = 1859 ±
565 25 Ma). Trace elements were measured using the same instrumentation as for U-Th-Pb, with the Attom
566 measuring in linkscan mode (see Supporting Information 3 for full analytical protocol), with
567 normalisation to GJ-1 zircon (Piazolo et al., 2017) except Ti which used Szymanowski et al. (2017).

568

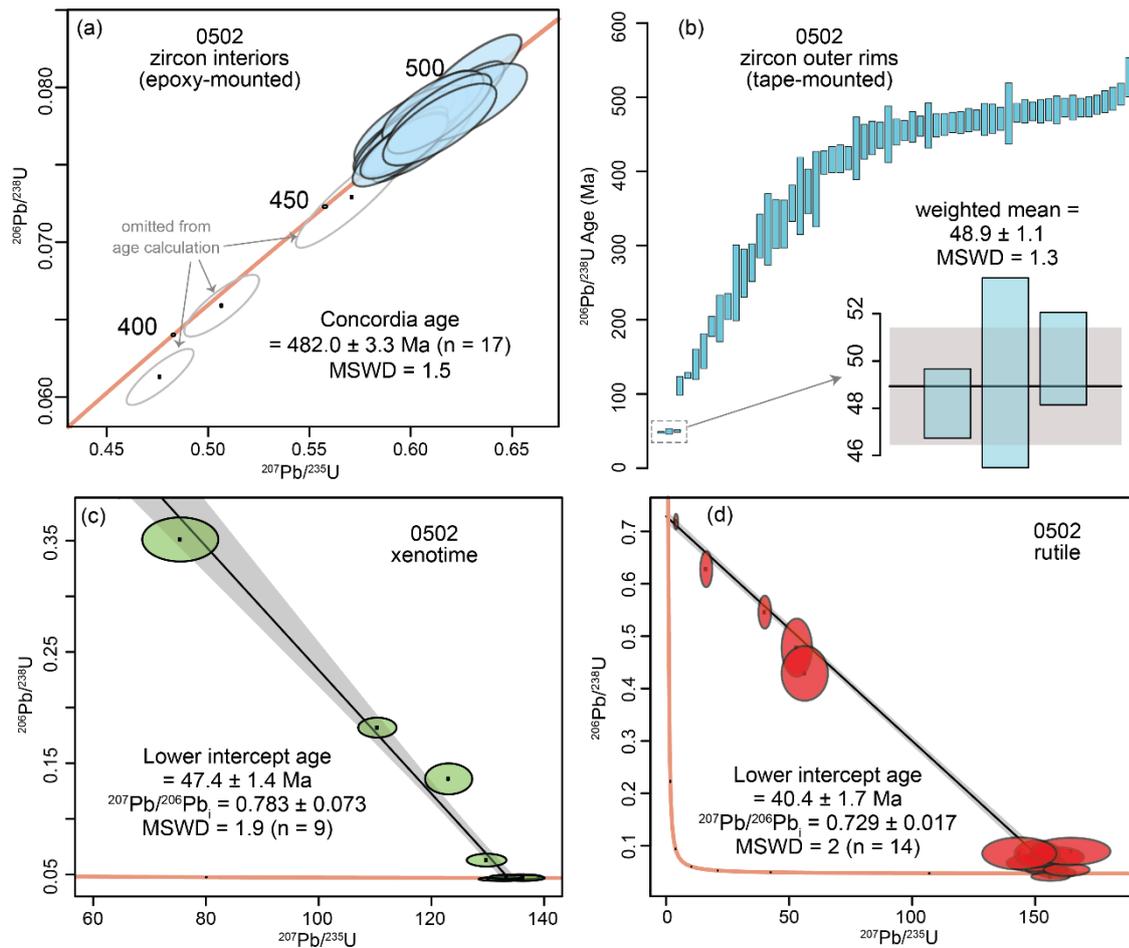
569 Zircon in one sample (a03-12) was further analysed by CA-ID-TIMS, following analytical and data
570 reduction methods described by Tapster et al. (2016), and utilising the ET535 EARTHTIME mixed
571 tracer (Condon et al., 2015). The CA-ID-TIMS uncertainties are quoted as analytical
572 only/analytical+tracer-calibration/analytical+tracer-calibration+decay-constant uncertainty, as per
573 community protocols. Our U-Pb dates can be compared using the intermediate uncertainty, but for the
574 sake of simplicity, we quote the final uncertainty within the discussion such that our dates can be
575 compared with constraints from other absolute and relative chronometry.

576

577 **4.2. Puga Gneiss (sample 05-02)**

578 **4.2.1. Zircon**

579 Zircon grains in the Puga gneiss sample 05-02 show oscillatory zoning and a euhedral shape, indicative
 580 of igneous zircon, with a Th/U of 0.05 – 0.26. Thin, bright rims (< 20 μm) are present surrounding the
 581 dark cores. From the epoxy mounted zircon, 33 LA-ICP-MS analyses of zircon cores were obtained
 582 from 21 grains. Five ages ranging from 2481 Ma to 1020 Ma (²⁰⁷Pb/²⁰⁶Pb age), indicate a population
 583 of xenocrystic zircon inherited during emplacement of the Puga gneiss igneous protolith. Nine analyses
 584 had ages that were discordant by >10 %, and were discarded from age calculation. Of the remaining 19
 585 concordant analyses, 17 spots provide a concordia age of 482 ± 5.9 Ma with a mean square of weighted
 586 deviates (MSWD) value of 1.5 (Fig. 6a). Using spot analysis of tape-mounted zircon, whereby the outer
 587 rim can be targeted more confidently, 64 grains yielded a spread of ages from ca. 569 Ma to 48 Ma
 588 (Fig. 6b). A broad plateau of ages overlaps the mounted zircon at ca. 480 Ma, consistent with our
 589 weighted mean concordia age. The youngest three analyses provide a weighted mean, using common
 590 lead corrected ²⁰⁶Pb/²³⁸U ages, of 48.9 ± 1.2 Ma (MSWD = 1.3; Fig. 6b).



591

592 Fig. 6. A. U-Pb Wetherill plot showing U-Pb zircon core analyses of Puga Gneiss sample 05-02. B.
 593 Rank-plot of all ²⁰⁶Pb/²³⁸U zircon dates. An older age plateau represents the zircon cores at ~480 Ma.
 594 The youngest dates represent thin rims, and converge on a Himalayan age with a weighted average

595 $^{206}\text{Pb}/^{238}\text{U}$ age of the three youngest zircon dates. C. Tera-Wasserburg plot of xenotime U-Pb analyses.
596 D. Tera-Wasserburg plot of rutile U-Pb analyses. Box heights and uncertainty ellipses are 2σ
597 uncertainties.

598

599 **4.2.2. Xenotime**

600 Rare 30-100 μm xenotime grains were observed in micaceous aggregates at corroded margins of garnet
601 in sample 05-02, as described above (Fig. 3a). The grains show homogenous brightness in BSE
602 indicating the absence of internal zonation. Nine spots were analysed across 9 grains, and yield a mixing
603 line between radiogenic and common lead components. Using a free regression, the lower intercept age
604 is calculated at 47.4 ± 1.4 Ma (MSWD = 1.9; Fig. 6c).

605

606 **4.2.3. Rutile**

607 Rutile grains in sample 05-02 measure 50-100 μm in size and show homogenous brightness in BSE,
608 indicating the absence of internal zonation. Fourteen spots were analysed across 14 rutile grains, and
609 yield a mixing line between radiogenic and common lead components. Using a free regression yields a
610 lower intercept age of 40.4 ± 1.7 Ma (MSWD = 2; Fig. 6d).

611

612 **4.3. Retrogressed mafic eclogite (sample a03-12)**

613 **4.3.1. Zircon**

614 Zircon grains are rarely found in mafic eclogites of the Tso Moriri Complex. In thin section, zircons
615 are often found as inclusions within eclogite-facies phases such as quartz and omphacite. Separated
616 zircon grains from sample a03-12 are translucent and colourless with rounded and euhedral grain shapes
617 and a grain size of < 70 μm (Fig. 7f). They have zoning patterns dominated by broad oscillatory or
618 sector zoning, that in some grains, truncates a darker core region; the latter were not analysed.

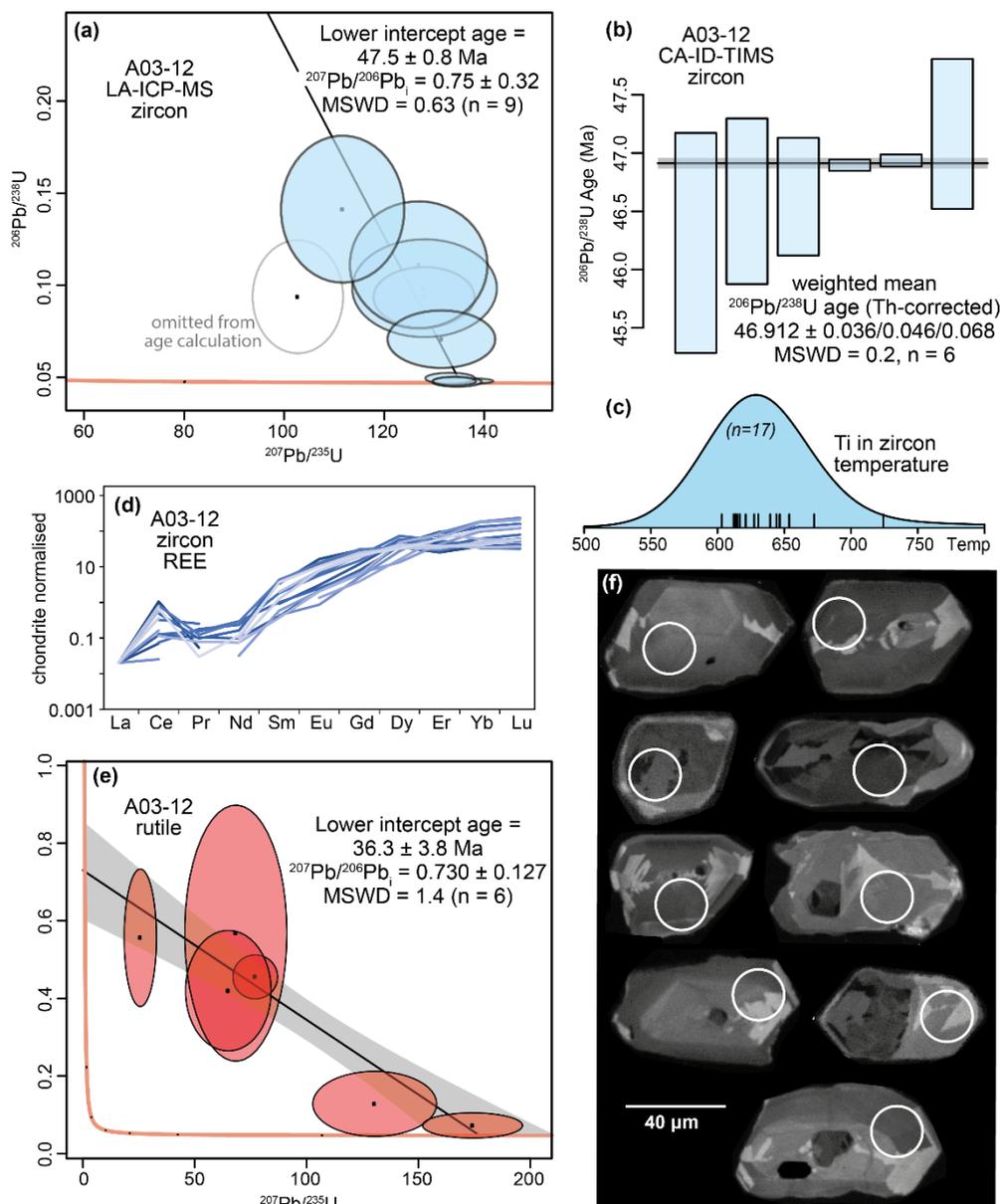
619

620 Thirteen laser spot analyses were performed on the cores of 13 zircon grains separated from sample
621 a03-12 for U-Pb, with 3 analyses rejected due to Pb counts below detection. The U concentrations are
622 variable (13.5 – 1528 ppm), and the degree of discordance varies between 0.1 and 12%. Of the
623 remaining 10 analyses, the lower intercept of 9 spots is calculated using a free regression at 47.5 ± 1.7
624 Ma (MSWD = 0.63; Fig. 7a). These zircon analyses have low Th/U ratios (< 0.01). The result implies
625 that the data conform to a single population. The omitted analysis has a much higher Th/U ratio (0.25),
626 suggesting an igneous core region was clipped during the ablation.

627

628 Trace elements were measured on 20 zircon grains, including adjacent spots on the same 13 grains
629 analysed for U-Pb; 3 analyses were omitted due to inclusions. Ti-in-zircon temperatures are calculated
630 using Si and Ti activities of 1.0, and the equation of Ferry and Watson (2007). The temperatures range

631 from 603 °C to 724 °C, forming a normal distribution around a peak at ca. 630 °C (Fig. 7c). The REE
 632 data are plotted as chondrite-normalised values (Fig. 7d). The REE patterns of zircon are broadly
 633 consistent across multiple grains, with no Eu anomaly, flat HREE patterns and depleted LREEs.
 634
 635 Several zircon grains were extracted from the resin mounts and prepared as single grain aliquots for
 636 CA-ID-TIMS. The resulting data are six reproducible fractions yielding a weighted mean (Th-
 637 corrected) $^{206}\text{Pb}/^{238}\text{U}$ age of 46.912 ± 0.068 Ma with an MSWD of 0.2 (Fig. 7b).
 638



639
 640 Figure 7. Mafic eclogite sample a03-12 a. Tera-Wasserburg plot of LA-ICP-MS U-Pb data for zircon,
 641 with lower intercept $^{206}\text{Pb}/^{238}\text{U}$ age. b. Weighted mean $^{206}\text{Pb}/^{238}\text{U}$ (Th corrected) age of 6 zircon grains
 642 using CA-ID-TIMS.uncertainties are quoted as analytical only/analytical+tracer-
 643 calibration/analytical+tracer-calibration+decay-constant uncertainty. c. Histogram of Ti in zircon

644 temperatures. d. Sample a03-12 zircon chondrite-normalised REE data. e. Tera-Wasserburg plot of LA-
645 ICP-MS U-Pb data for rutile, with lower intercept $^{206}\text{Pb}/^{238}\text{U}$ age. f. Cathodoluminescence images of
646 zircon separates from sample a03-12. All box heights and uncertainty ellipses are 2σ uncertainties.

647

648 **4.3.2. Rutile**

649 Rutile grains in sample a03-12 show homogenous brightness in BSE with no indication of internal
650 zonation. The majority of analyses had Pb counts below detection, the remaining data comprise six
651 spots analysed across 6 grains. The analyses are distributed between the radiogenic and common lead
652 components, and using a free regression, yield a lower intercept age of 36.3 ± 3.8 (MSWD = 1.4; Fig.
653 7e).

654

655 **4.4. Summary and Interpretations**

656 **4.4.1. Puga Gneiss (sample 05-02)**

657 Zircon core ages of 482.0 ± 5.9 Ma in the Puga Gneiss are interpreted to represent the igneous
658 crystallisation age and are comparable with U-Pb zircon ages of 479 ± 1 Ma in both the Polokongka La
659 granite and the Puga Gneiss (Girard and Bussy, 1999). A monazite age of 473 ± 9 Ma in the Polokongka
660 La granite, interpreted to have formed during crystallisation of the granite (Bidgood et al., 2022), also
661 overlaps with the age of zircon crystallisation. Rim analyses of Ordovician zircon from the Puga Gneiss
662 yielded an age of 48.9 ± 1.2 Ma (LA-ICP-MS), which overlap with high precision CA-ID-TIMS zircon
663 age from mafic eclogite a03-12 (hosted within the gneiss), suggesting that zircon rims in the Puga gneiss
664 also grew at eclogite-facies conditions.

665

666 Xenotime is found exclusively in Puga Gneiss sample 05-02, in mica aggregates adjacent to the
667 corroded outer rim of a large garnet porphyroblast (Figs. 2 and 3). Breakdown of parts of the Mg-
668 enriched garnet rims, formed under peak eclogite-facies conditions (M2), liberated Y, which is
669 concentrated in garnet zones P2 and P3 (Fig. 3a), for xenotime growth. We therefore interpret the U-Pb
670 xenotime age in sample 05-02 (47.4 ± 1.4 Ma) as the age of initial garnet breakdown during
671 decompression. The xenotime age overlaps with the eclogite facies zircon rim age within the same
672 sample 05-02 (48.9 ± 1.2 Ma), as well as the precise eclogite-facies zircon CA-ID-TIMS age of mafic
673 eclogite sample a03-12 (46.912 ± 0.068 Ma), located within the same structural unit. This allows us to
674 assign the xenotime crystallization and garnet breakdown in the Puga Gneiss sample to the M3a
675 metamorphic assemblage, reflecting the earliest phase of decompression at eclogite facies conditions.

676

677 Rutile is present as inclusions within prograde (M1) to peak pressure (M2) garnets and the matrix of
678 the Puga Gneiss. Given that the peak temperature of the Tso Morari Complex ($705\text{-}755^\circ\text{C}$, St-Onge et
679 al., 2013) is greater than the closure temperature of rutile at $\sim 400\text{ - }630^\circ\text{C}$ (Cherniak, 2000; Koojiman
680 et al., 2010; Mezger et al., 1989; Vry & Baker, 2006), which is dependent on cooling rate and grain size

681 (Zack and Koojiman, 2017; Oriolo et al., 2018), the U-Pb age of rutile can therefore be attributed to
682 cooling of the Puga Gneiss through the rutile closure temperature at 40.4 ± 1.1 Ma, after peak
683 temperature conditions.

684

685 **4.4.2. Retrogressed mafic eclogite (sample a03-12)**

686 LA-ICP-MS analyses of zircon from mafic eclogite sample a03-12 are entirely <50 Ma, and provide a
687 weighted mean age of 47.5 ± 1.7 Ma. Although a limited dataset (9 analyses), our result implies a single
688 population of metamorphic zircon, rather than preserving a protracted history of zircon growth. Flat
689 HREE profiles and the lack of an Eu anomaly in the core and mantle of zircon grains are indicative of
690 crystallisation in garnet-present, plagioclase-absent conditions, consistent with crystallisation in the
691 eclogite facies (Schaltegger et al., 1999; Rubatto, 2002; Rubatto and Hermann, 2003). The measured
692 zircon date is therefore interpreted to correspond to growth within the eclogite-facies. Ti in zircon
693 temperatures of 603 °C to 724 °C lie below the closer temperature of Ti in zircon (~ 900 °C, Watson et
694 al., 2006; Cherniak et al., 2007) and is therefore interpreted as the zircon crystallisation temperature.

695

696 The zircon CA-ID-TIMS data from mafic eclogite sample a03-12 are tightly clustered yielding a date
697 of 46.91 ± 0.068 Ma, which we interpret as the age of zircon growth at eclogite facies conditions (M1-
698 3a). We note that this interpretation is heavily weighted to only the two analyses of the most U-rich
699 zircon, excluding these analyses yields a weighted mean of $46.71 \pm 0.33/0.38/0.38$ Ma (MSWD = 1.04;
700 $n=4$) and is therefore within uncertainty of the original interpretation. Regardless of interpretation, the
701 uncertainty associated with the CA-ID-TIMS data is 1 to 2 orders of magnitude smaller than LA-ICPMS
702 U-Pb data and provides the most precise estimate of the timing of metamorphic zircon growth from Tso
703 Morari thus far. Additionally, even with the improved precision, the analyses and the lack of dispersion
704 indicated by their MSWDs suggests no measurable crystal to crystal variation at around the ~ 1 Myr
705 resolution.

706

707 Rutile is the peak titanium-bearing phase in mafic rocks during eclogite-facies metamorphism and is
708 present as inclusions in garnet (M1), as well as in the matrix (M2-3b). Peak temperature in the Tso
709 Morari Complex is estimated at $600 - 755$ °C (De Sigoyer et al., 2000; St-Onge et al., 2013), above the
710 predicted closure temperature of rutile at $\sim 400 - 630$ °C (Cherniak, 2000; Koojiman et al., 2010;
711 Mezger et al., 1989; Vry & Baker, 2006) which is dependent on cooling rate and grain size (Zack and
712 Koojiman, 2017; Oriolo et al., 2018). The U-Pb age of rutile can therefore be attributed to cooling
713 through this closure temperature range.

714

715

716 **5. Metamorphism, deformation and geochronology of the Tso Morari** 717 **Complex**

718

719 Previous studies indicate that subduction, exhumation and emplacement of the Tso Morari Complex
720 took place between c. 60 Ma and c. 7.5 Ma (Fig. 6), with exhumation to lower crustal conditions by
721 45.3 ± 1.1 Ma at an average rate of ~ 12 mm a⁻¹ (St-Onge et al., 2013). The texture and composition of
722 the mafic eclogites and Puga gneiss samples in this study collectively provide a record of initial
723 exhumation from UHP eclogite-facies conditions, followed by exhumation through crustal conditions.
724 Integrating this information with our new high-precision geochronology from a range of accessory
725 phases which crystallized at different stages of metamorphism allows us to constrain the timing of
726 mineral growth and fabric development with respect to the burial and exhumation of the Tso Morari
727 Complex.

728

729 Evidence of prograde metamorphism to peak pressures (M1-2) is preserved within garnets from mafic
730 eclogite and the Puga Gneiss, comparable to that observed in the eclogite studied by St-Onge et al
731 (2013). The garnet-bearing Puga Gneiss, as represented by sample 05-02, is not typical of the host Puga
732 Gneiss, which generally preserves only amphibolite-facies assemblages. The formation and
733 preservation of garnet in this sample is therefore particularly valuable given that it records the same
734 overall prograde compositional trends as seen in the Tso Morari eclogites. The analysed garnet (Fig.
735 3a,b) also contains complex deformation fabrics and inclusion suites, indicating that deformation and
736 transformation of the original granite was already underway prior to garnet growth (M1) in at least part
737 of the complex. Early fabrics are rarely preserved in the Tso Morari Complex, where the earliest stage
738 of macroscopic deformation has been previously identified as the dominant top-to-the-east exhumation
739 fabric within the Puga Gneiss, and is attributed to initial M3a exhumation from eclogite-facies
740 conditions (Epard and Steck, 2008).

741 Our eclogite-facies zircon dates of 46.912 ± 0.068 Ma (CA-ID-TIMS) and 47.5 ± 1.7 Ma (LA-ICPMS)
742 from a mafic eclogite overlap with our zircon rim and xenotime dates in the Puga Gneiss. We interpret
743 this overlap as a record of the earliest phase of decompression at 630 °C and eclogite facies conditions,
744 as recorded by the Ti in zircon temperatures and the partial breakdown of yttrium-bearing garnet rims.
745 The crucial breakdown of the garnet rim zones and fracture-based alteration, with xenotime among its
746 products, provides insight into the early exhumation history of these rocks that is not recorded
747 elsewhere, but predates the Barrovian overprint of 7.0 - 8.4 kbar, 705 - 755 °C recorded at 45.3 ± 1.1
748 Ma and 43.3 ± 1.1 Ma by St-Onge et al. (2013).

749

750 The formation of quartz microstructures during high-temperature ($> 530^{\circ}\text{C}$) dynamic recrystallisation
751 indicates that deformation took place during exhumation to crustal conditions (M3b). There is no
752 significant later overprinting of quartz deformation fabrics in Puga Gneiss samples from the core of the
753 dome (Bidgood, 2020), implying a lack of pervasive deformation below 530°C . Deformation therefore
754 occurred prior to cooling through the rutile closure temperature of $\sim 630\text{-}400^{\circ}\text{C}$ (Cherniak, 2000; Li et
755 al., 2013; Mezger et al., 1989; Vry & Baker, 2006) at 40.4 ± 1.1 Ma and was ductile and pervasive. This
756 interpretation is confirmed by Dutta and Mukherjee (2021) who predict deformation temperatures $>$
757 600°C based on misorientation analysis and micro-textural evidences of incipient partial melting in the
758 Puga gneiss. Subsequent deformation during exhumation relating to doming and emplacement (M3c)
759 was not pervasive, with foliations and lineations developing at the margins of the dome, adjacent to the
760 normal sense shear zones (e.g. Zildat-Ribil fault, Epard and Steck, 2008; Bidgood, 2020; Dutta and
761 Mukherjee, 2021). The age of exhumation of the Tso Morari dome is recorded by the Ar-Ar muscovite
762 and biotite and apatite fission track dates of < 32.4 Ma (De Sigoyer et al., 2000; Schlup and Carter,
763 2003) which record cooling through the Ar closure temperatures of muscovite ($\sim 425^{\circ}\text{C}$, Harrison et
764 al., 2009), biotite ($< 345^{\circ}\text{C}$, Harrison et al., 1985) and apatite ($< 135^{\circ}\text{C}$, Warnock et al., 1997)
765 respectively.

766

767 **6. Discussion**

768 **6.1. Continental subduction and exhumation in the NW Himalaya**

769 Our petrographic correlation of zircon and xenotime ages with the M3a assemblage indicates that zircon
770 growth at $\sim 47\text{-}46$ Ma took place at subsolidus conditions during the earliest stages of decompression
771 from UHP conditions. These ages overlap with the zircon age distribution peak of $47\text{-}43$ Ma recorded
772 by Donaldson et al. (2013). Considering their analytical scatter (i.e. MSWDs of 2.4 and 3.4) the
773 Donaldson et al. dates have reasonable agreement with our $47\text{-}46$ Ma age, however, those data were
774 previously interpreted as a record of UHP metamorphism, starting at ~ 47 Ma. It is therefore necessary
775 to reassess the Donaldson et al. (2013) data in light of our new data and observations, as follows.

776 Subsolidus zircon growth requires a fluid phase to mediate the liberation of Zr from Zr-bearing phases
777 (ilm, ru, cpx, grt) (Chen et al., 2010; Kohn et al., 2015; Chen and Zhang, 2017; Skuzovatov et al., 2021).
778 The host rock in the Tso Morari complex is a dry granite with intruded mafic dykes (see Bidgood et al.,
779 2023) resulting in restricted opportunity for prograde/peak zircon growth. Palin et al. (2014) determined
780 that the first pulse of post-peak fluid in the eclogite facies relates to the destabilisation of talc and growth
781 of the coarse-grained amphiboles at ~ 23 kbar, followed by fluid infiltration from an external source at
782 ~ 19 kbar. Coarse-grained, zoned amphiboles are abundant in mafic eclogites from across the Tso
783 Morari Complex suggesting that post-peak eclogite facies hydration was a common and widespread
784 occurrence, aided further by exhumation-related deformation. These influxes of fluid during
785 exhumation, subsequent to a potentially dry spell at prograde to peak conditions, would have provided

786 conditions favourable for high concentrations of zircon growth. Consequently, we argue that the
787 breakdown of UHP garnet rims recorded by xenotime at 47.4 ± 1.4 Ma indicates that the zircon age
788 peak at 47 – 43 Ma of Donaldson et al. (2013), along with our own CA-ID-TIMS zircon age of $46.91 \pm$
789 0.068 Ma, reflects a pulse of zircon growth during the onset of exhumation from UHP conditions, aided
790 by the exhumation-driven liberation of fluids.

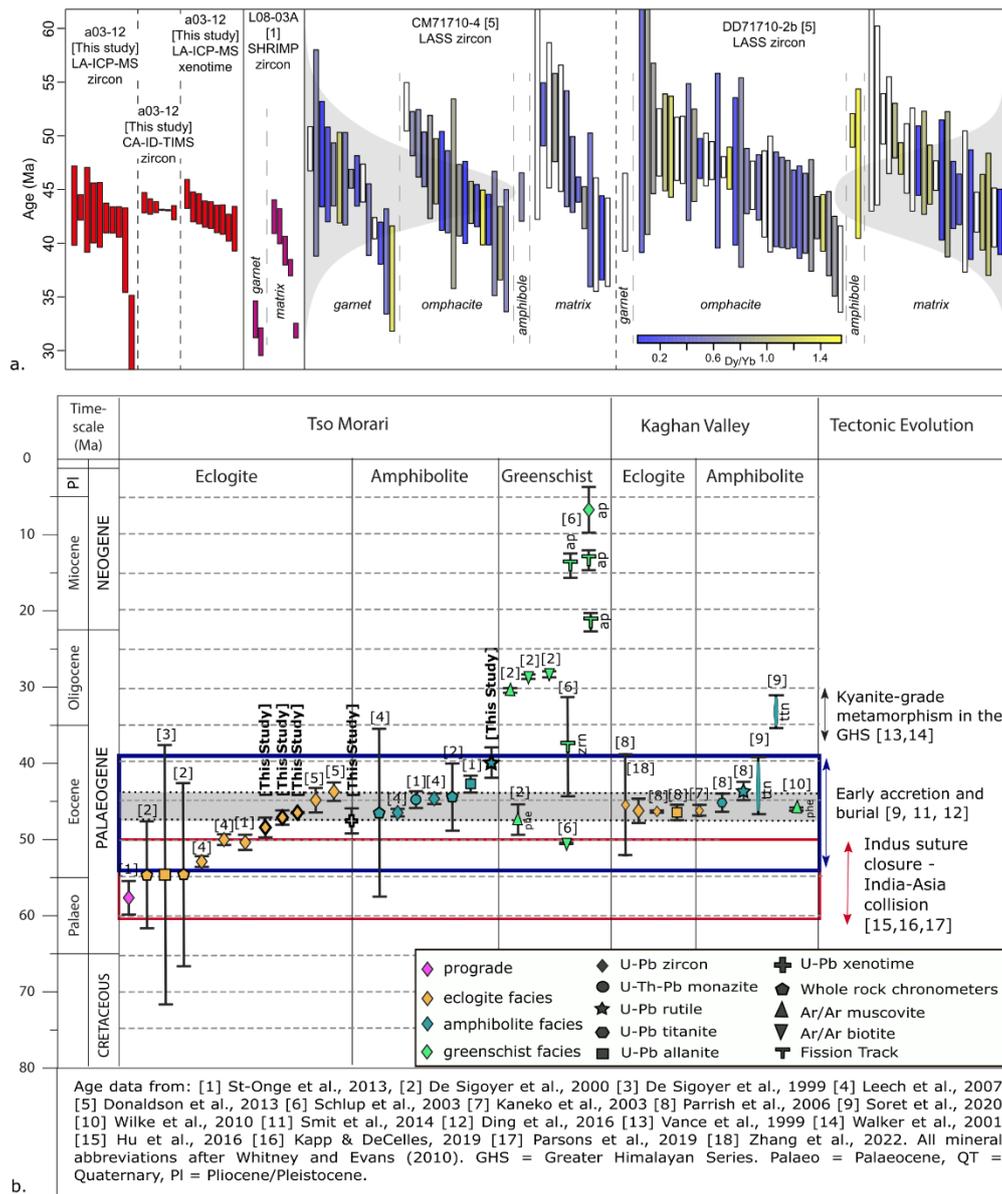
791 Based on the above conclusion, older dates in previous studies, including a 58 ± 2.2 Ma ($n = 2$) zircon
792 age from St-Onge et al (2013) and apparent older zircon dates (55 – 48 Ma) from Donaldson et al.
793 (2013), may record zircon growth during prograde to peak metamorphism. However, it is difficult to
794 place them in a robust context given: 1) the proportion of common lead for most of the analyses; 2) the
795 potential ablation of older zircon domains reflecting the protolith age; and 3) the lack of relationship
796 between textural location and REE chemistry (Fig. 8a). The nature of apparent younger common-lead
797 corrected dates from the Donaldson et al. (2013) dataset also remain uncertain for the same reasons;
798 younger dates could plausibly record continued zircon growth during decompression, but dates younger
799 than the ca. 40 Ma rutile age imply these may relate to analytical inaccuracies. Given that we consider
800 our xenotime date as a marker for the onset of UHP exhumation, we suggest that the Donaldson et al.
801 (2013) data reflect zircon growth due to fluid infiltration at and after 47 Ma.

802 Our new CA-ID-TIMS zircon age of 46.912 ± 0.068 Ma more closely and precisely correlates with ages
803 from zircon (46.4 ± 0.1 Ma – ID-TIMS; 46.2 ± 0.7 – SHRIMP, 46 ± 2 Ma - SIMS) and allanite (46.5
804 ± 1.0 Ma – ID-TIMS) in UHP assemblages from Kaghan, located 450 km to the west of Tso Morari
805 (Kaneko et al., 2003; Parrish et al., 2006; Zhang et al., 2022) (see Fig. 8). We do not think this is
806 coincidental; metamorphic P-T data from both of these units record similar prograde (up to UHP
807 metamorphism) and retrograde (including UHP exhumation and exhumation through the crust) P-T
808 paths. In Kaghan, coesite is found in thin metamorphic zircon rims in the felsic gneiss, indicating that
809 zircon crystallisation occurred at UHP conditions (Kaneko et al., 2003). In the Tso Morari Complex,
810 coesite is observed in the outermost rims of prograde garnets but has not yet been observed as inclusions
811 in zircon. Quartz microstructures after the presence of former coesite are also reported from the
812 Polokongka La Granite of the Tso Morari Complex (Bidgood et al., 2021). Based on these similarities,
813 we argue that the overlap in ages between Tso Morari and Kaghan indicates that UHP exhumation and
814 associated fluid flux at 47- 46 Ma was responsible for a ubiquitous pulse of zircon growth across the
815 NW Himalaya, occurring at the same time as or shortly after the achievement of maximum P and T
816 under UHP conditions. We suggest that the regional synchronicity between Tso Morari and Kaghan,
817 across a distance of ~450 km, reflects the scale at which slab dynamics control metamorphism and
818 exhumation with a subduction zone setting.

819 The mode of UHP exhumation is unclear from our data. Isothermal, triclinic deformation during
820 exhumation of the Tso Morari complex, as reported by Long et al. (2020) and Dutta and Mukherjee

821 (2021), is most compatible with the recirculation model and plunger model of Warren et al. (2008a,
822 2008b, 2008c). These models invoke the transport of crustal slices of the lower plate from UHP
823 pressures along the subduction interface, rather than wholesale exhumation of the subducting slab, and
824 can therefore occur independently from, and without need for slab break-off. The timing of slab break-
825 off cannot be resolved from our data and therefore remains debated, with estimates ranging widely
826 between ~50-10 Ma (see reviewed compilation of Garzanti et al., 2018).

827 Lastly, cooling through the rutile closure temperature took place ~ 3.7 Ma later in Tso Morari than
828 Kaghan (see Fig. 8). At face value, this implies a longer period of time between zircon growth and
829 exhumation through the rutile closure temperature for the Tso Morari Complex, relative to Kaghan.
830 However, uncertainties in the exact conditions of zircon growth and rutile closure temperature prevents
831 us from making any meaningful interpretation for the cause of this difference.



832

833 Figure 8: a. Zircon data for individual samples from this study, St-Onge et al., (2013) and Donaldson et al., (2013). Blue-yellow colour scale represents Dy/Yb. Solid colours do not contain individual REE
 834 al., (2013). Blue-yellow colour scale represents Dy/Yb. Solid colours do not contain individual REE
 835 spot data. Sample histograms shown in grey for Donaldson et al., (2013). b. Time chart for Tso Morari
 836 and Kaghan, after (Palin et al., 2012), comprising data from multiple sources, including this study.

837

838 6.2. Geodynamic significance of UHP exhumation during the Himalayan orogeny

839

840 Our data suggest that zircon crystallisation recorded in the Tso Morari complex and Kaghan at 47-46
 841 Ma, corresponds to the onset of exhumation of Indian continental crust from UHP conditions. In the
 842 following sections, we consider the geodynamic significance of this event with respect to the wider

843 metamorphic, magmatic and plate kinematic evolution of the Himalayan orogeny and the India-
844 Australia-Eurasia plate network (Figs. 9 & 10).

845 **6.2.1. UHP exhumation during the Himalayan orogeny**

846 When placed in context with the bedrock record of collision, our constraints for the onset of UHP
847 exhumation in the NW Himalayan reveal how changes in subduction dynamics affected the structural,
848 magmatic, and metamorphic evolution of the Himalayan orogeny (Fig. 9).

849 In the lower plate (Indian crust), the onset of UHP exhumation at 47-46 Ma overlaps with the onset of
850 local prograde amphibolite facies, Barrovian-style metamorphism in the NW Himalaya at 47 Ma to 39
851 Ma (Soret et al., 2021) (Fig. 9a). This was followed by wide-spread amphibolite facies metamorphism
852 of the Himalayan metamorphic core (HMC) across the rest of the Himalaya during crustal thickening
853 of the HMC from ~41 Ma through to ~17 Ma (Fig. 9a) (Ambrose et al., 2015, Carosi et al., 2016,
854 Goscombe et al., 2018, Carosi et al., 2019, Mottram et al., 2019, Waters, 2019; Ji et al., 2024), whilst
855 high-pressure eclogite to granulite facies metamorphism of Indian lower crust is recorded in localized
856 occurrences across the orogen from 40 Ma to 25 Ma (Fig. 9a) (O'Brien, 2019a, Chen et al., 2022).

857 UHP exhumation at 47-46 Ma also overlaps with significant magmatic and metamorphic changes in the
858 upper plate of the orogen (Eurasian crust)(Fig. 9a; see also Fig. 10c for locations). Between 50 Ma and
859 40 Ma, isotopic signatures of magmatic rocks from the Kohistan-Ladakh batholith record increased
860 crustal contamination of arc magmatism (Fig. 9a) (Bouilhol et al., 2013, Jagoutz et al., 2019). Along
861 the Lhasa block, the Gangdese arc records adakite magmatism from ~48 Ma produced by melting of
862 the Tibetan lower crust (e.g., Searle et al., 2011, Guan et al., 2012, Ma et al., 2014), and the cessation
863 of subduction-related magmatism by ~40 Ma (Fig. 9a)(e.g., Zhu et al., 2019). At the same time, the
864 Lhasa block recorded high pressure-low temperature kyanite-grade partial melting associated with
865 deformation and crustal thickening at 44-32 Ma (Fig. 9a) (Zhang et al., 2010, Palin et al., 2014). Further
866 to the east, the Mogok metamorphic belt and Eastern Ophiolite Belt in Myanmar record sillimanite-
867 grade metamorphism between 48-22 Ma, which included a phase of granulite facies metamorphism
868 between 43-32 Ma (Fig. 9a) (Barley et al., 2003, Searle et al., 2007, Searle et al., 2017, Searle et al.,
869 2020, Lamont et al., 2021).

870 Collectively the structural, metamorphic, and magmatic changes recorded in both the lower and upper
871 plates of the Himalayan orogeny between ~50-40 Ma (Fig. 9a) reflect (1) a warming metamorphic
872 thermal gradient (e.g., Soret et al., 2021, Chen et al., 2022; Ji et al., 2024); (2) greater mechanical
873 coupling of the upper and lower plates resulting in increased crustal shortening and thickening; and (3)
874 a greater contribution of Indian continental crustal material to upper plate magmatism. These
875 phenomena are best explained by a reduction in the dip of the subducting Indian plate beneath Eurasia
876 during that time (Fig. 10a-b) (e.g., Soret et al., 2021, Chen et al., 2022; Ji et al., 2024). We attribute this
877 to the early stages of the India-Asia collision *sensu stricto* (i.e., second collision of the Himalayan

878 orogeny), whereby the positive buoyancy of the Indian continental crust stalled subduction of the Indian
879 plate beneath the Eurasian margin (Fig. 9b).

880 **6.2.2. A shift in orogenic regimes**

881 UHP exhumation at 47-46 Ma marks a significant geodynamic shift in the orogenic regime of the
882 collision zone (Figs. 9b, 10a,b). Prior to this time, India-Asia convergence was accommodated by
883 subduction of Indian and/or Neotethys oceanic lithosphere beneath Eurasian. Then, following the India-
884 Asia continent-continent collision *sensu stricto* by 47 Ma, the reduced dip of the Indian slab increased
885 the mechanical coupling between upper and lower plates and the component of convergence
886 accommodated by crustal shortening. This switch in mode of convergence can be viewed as a shift in
887 the orogenic regime of the collision zone (cf. Cawood et al., 2009) from (1) an *accretionary orogen*, in
888 which India-Asia convergence was driven and accommodated by subduction of Indian and/or Neotethys
889 oceanic lithosphere (Fig. 10a); to (2) a *collisional orogen*, in which India-Asia convergence was
890 accommodated by crustal shortening (e.g., Replumaz et al., 2014; Parsons et al., 2021; Chen et al.,
891 2022), and driven by northeastward subduction of Indian-Australian oceanic lithosphere beneath SE
892 Asia, further to the east (Fig. 10b) (e.g., Li et al., 2008; Capitanio et al., 2015; Parsons et al., 2021, Bose
893 et al., 2023). The onset of UHP exhumation at 47-46 Ma therefore timestamps the change in orogenic
894 regime of the India-Asia collision zone from an *accretionary orogen* to a *collision orogen* (cf. Cawood
895 et al., 2009) (Fig. 9b).

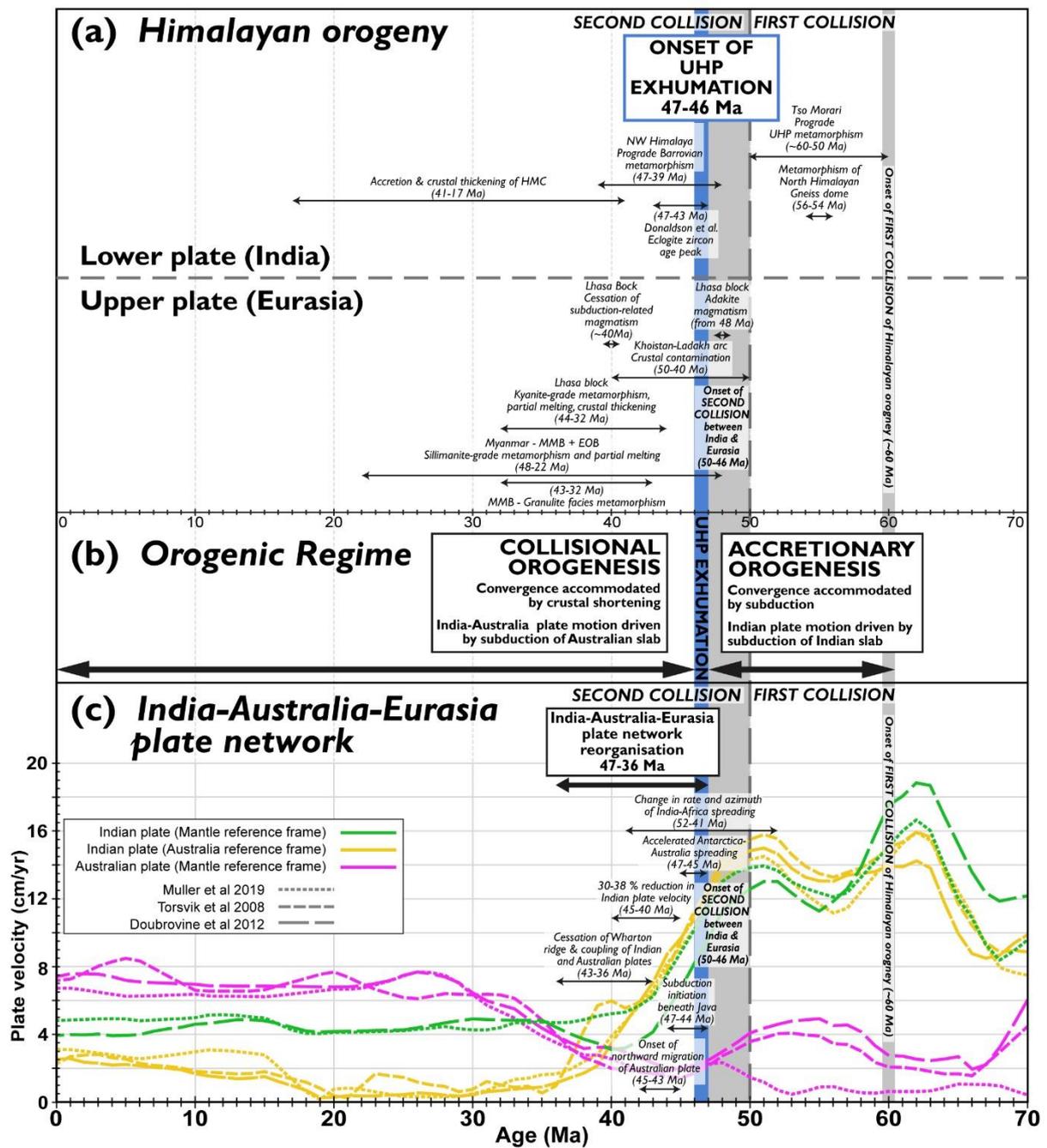
896 **6.3.2. Implications for models of the Himalayan orogeny and the India-Asia continent-continent** 897 ***collision (sensu stricto)***

898 We interpret the onset of the India-Asia collision (*sensu stricto*) to occur at ~50-47 Ma, based on (1)
899 our constraint for the onset of UHP exhumation at 47-46 Ma; (2) a reduction in Indian plate velocity
900 starting at ~50 Ma (Fig. 9c); (3) plate reconstructions of the restored Indian and Eurasian margins at
901 50-45 Ma (Fig. 10b) (Replumaz et al., 2014; Parsons et al 2021); and (4) changes in the magmatic and
902 metamorphic evolution of the Eurasian margin (upper plate) between 50-40 Ma (listed above)(Fig. 9a).

903 In the context of our data, single collision models initiating at ~60 Ma (e.g., Hu et al., 2016, Ingalls et
904 al., 2016) would require ~14 Myrs of uninterrupted continental subduction prior to UHP exhumation of
905 Tso Morari at 47-46 Ma. This equates to wholesale subduction of a 2500 x 2000 km area of Indian
906 continental lithosphere. This scenario remains highly disputed and is not discussed further as it is yet to
907 be shown that such extreme volumes of continental subduction are feasible or sustainable (Afonso and
908 Zlotnik, 2011, Ingalls et al., 2016, van Hinsbergen et al., 2019, Parsons et al., 2020).

909 In the context of double-collision models for the Himalayan orogeny, prograde to peak metamorphism
910 of the Tso Morari complex between 60-50 Ma (Leech et al., 2007; St-Onge et al., 2013; Donaldson et
911 al., 2013), as well as Barrovian metamorphism of the North Himalayan gneiss domes between 56-54

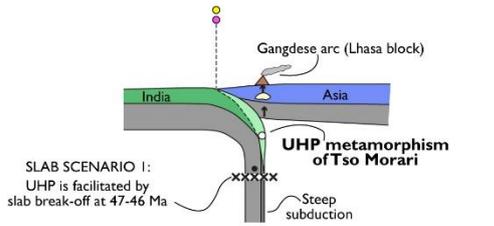
912 Ma (Smit et al., 2014, Ding et al., 2016), may be explained by initial burial of the NW Himalaya during
913 the first collision event of the Himalayan orogeny (Fig. 9a) (first collision began at ~61 Ma, see
914 introduction for definitions). Our new data do not permit us to constrain the validity of the opposing
915 hypotheses for first collision beyond that which is already discussed in recent reviews (e.g., Kapp &
916 DeCelles, 2019; van Hinsbergen et al., 2019; Parsons et al., 2020). However, we note that models
917 proposing first collision of the Himalayan orogeny between India and an equatorial intra-oceanic arc at
918 ~61 Ma require two collisions at two separate subduction zones (an intra-oceanic subduction zone and
919 a continental margin subduction zone) and therefore require two slab break-off events. We are unaware
920 of any studies (including our own), which present bedrock evidence for more than one break-off event
921 during the Himalayan orogeny.



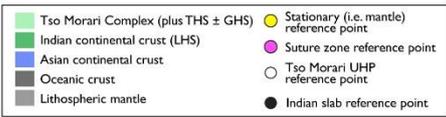
922

923 Figure 9. UHP exhumation at 47-46 Ma and its temporal relationship with the metamorphic, magmatic,
 924 and plate kinematic evolution of the Himalayan orogeny and India-Australia-Eurasia plate network. (a)
 925 Metamorphic and magmatic events in the Himalayan orogeny. (b) Orogenic regime of the Himalayan
 926 orogeny: onset of UHP exhumation marks the transition from an accretionary orogen to a collisional
 927 orogen (c.f., Cawood et al., 2009). (c) Reorganization of the India-Australia-Eurasia plate network at
 928 47-36 Ma with plate velocity profiles for the Indian and Australian plates (Torsvik et al., 2008,
 929 Doubrovine et al., 2012, Müller et al., 2019) plotted using GPlates (Müller et al., 2018). Data sources
 930 for events in (a) and (c) are cited in the main text. (c) is modified after Parsons et al. (2021). EOB –
 931 Eastern Ophiolite Belt; HMC – Himalayan Metamorphic core; MMB – Mogok Metamorphic Belt.

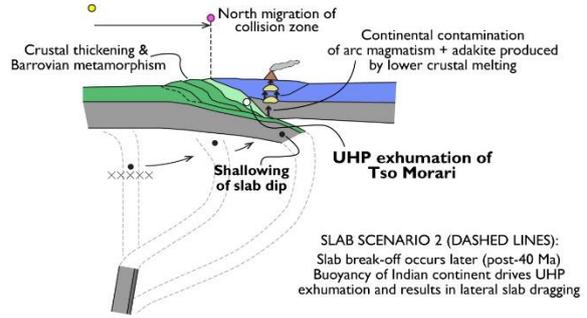
(a) 50-47 Ma: Onset of India-Asia collision



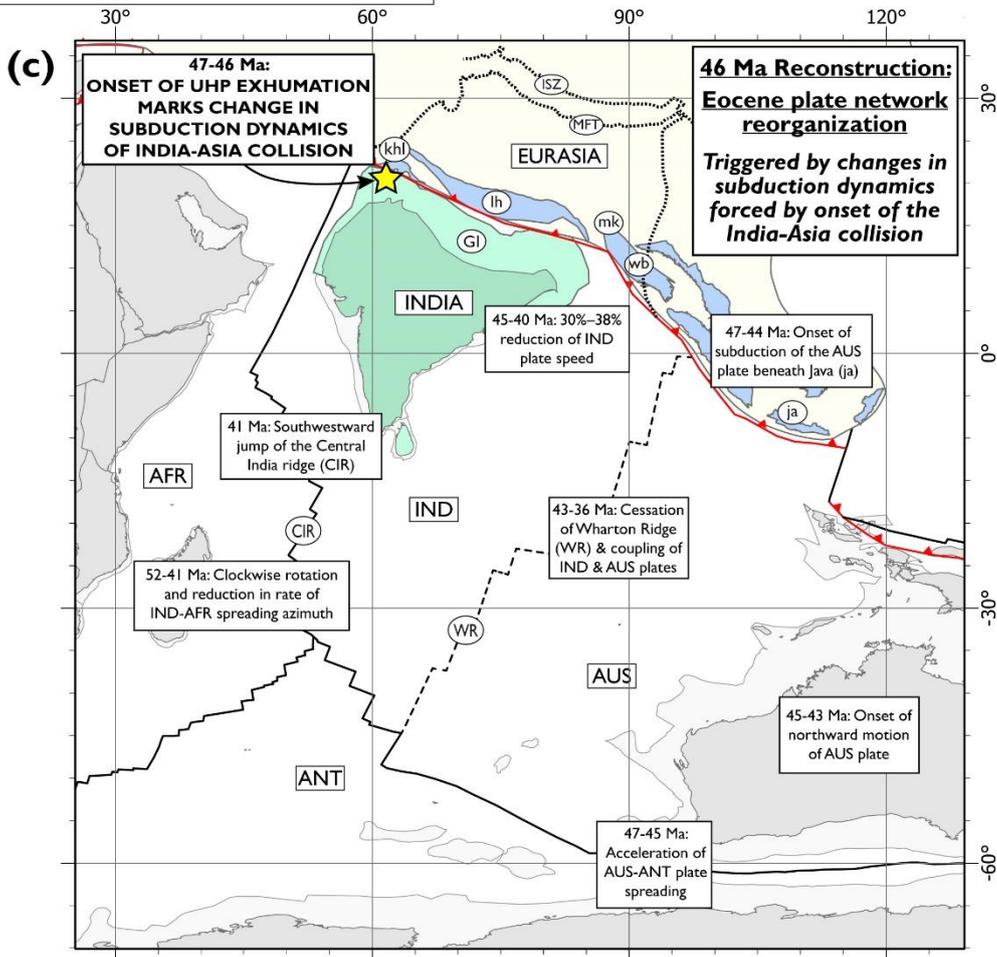
Prior to collision:
Asian active margin = Accretionary Orogen
Convergence accommodated by subduction
Convergence driven by subduction of Indian slab



(b) 47-46 Ma and younger



India-Asia collision zone = Collisional Orogen
Convergence accommodated by crustal shortening ± continental subduction
Convergence driven by subduction of Australian slab



932

933 Figure 10. The geodynamic significance of UHP exhumation for orogenesis and plate tectonics. (a-b)
 934 Schematic cartoon section of the India-Asia collision showing the postulated shallowing of Indian slab
 935 dip (e.g., Chen et al 2022), marked by the onset of UHP exhumation at 47-46 Ma (not to scale). (a)
 936 Steep subduction and UHP metamorphism during onset of the India-Asia collision (~50-47 Ma). Prior
 937 to collision, the Asian active continental margin represented an Accretionary Orogen (c.f. Cawood et
 938 al., 2009) in which convergence was driven and accommodated by subduction of the Indian and/or
 939 Neotethys slab. (b) UHP exhumation and slab shallowing starting at 47-46 Ma, in response to the

940 positive buoyancy of Indian continental crust following the onset of continental collision. The impacts
941 of the reduced slab dip are annotated on (b), showing an increase in crustal shortening, Barrovian
942 metamorphism, crustal contamination of arc magmatism and lower crustal melting of the Gangdese arc
943 to produce adakite. The collision zone represents a Collisional Orogen (c.f. Cawood et al., 2009),
944 whereby convergence is accommodated by crustal shortening \pm continental subduction and driven by
945 subduction of the Australian slab to the east (see discussion). Two alternative slab break-off scenarios
946 are presented. Slab Scenario 1 (presented in (a)): UHP exhumation and slab dip shallowing occurred in
947 response to slab break-off at 47-46 Ma. Slab Scenario 2 (presented in (b)): Alternatively, UHP
948 exhumation occurred via detachment of the Tso Moriri complex from the Indian slab, whilst the
949 buoyancy of the Indian continental crust resulted in lateral slab dragging, overturning of the trailing
950 oceanic slab and reduction in slab dip (see Parsons et al., 2021 for further discussion of lateral slab
951 migrations during the India-Asia collision), with slab break-off occurring at a later time (see discussion).
952 (c) 46 Ma plate reconstruction showing tectonic events of the Eocene plate network reorganization,
953 triggered by changes in the subduction dynamics of the India-Asia collision. UHP exhumation at 47-46
954 Ma provide a precise timestamp linking changes in the subduction dynamics of the India-Asia collision
955 zone to plate kinematic changes of the wider plate network. Reconstruction drawn from Müller et al.
956 (2019) using GPlates (Müller et al., 2018), with Greater India and Eurasian margin subduction zone
957 outlines constrained by Parsons et al. (2020; 2021). Thick dashed black lines show present day locations
958 of the Indus Suture Zone (ISZ) and Main Frontal Thrust (MFT) of the Himalayan orogen, , requiring
959 1000-2000 km of crustal shortening since ~47-46 Ma (Parsons et al., 2021). Abbreviations: AFR –
960 African plate; ANT – Antarctic plate; AUS – Australian plate; CIR – Central Indian ridge; GI – Greater
961 India; IND – Indian plate; ISZ – Indus Suture Zone; ja - Java; khl – Khoistan-Ladakh block; lh – Lhasa
962 block; MFT – Main Frontal Thrust; mk – Mogok metamorphic belt; wb - West Burma block; WR –
963 Wharton ridge

964 **6.2.4. UHP exhumation during Eocene plate network reorganization**

965 From a plate-kinematic perspective (Fig. 10c), UHP exhumation at 47-46 Ma coincides with a
966 significant reorganisation of the India-Eurasia-Australia plate network during the mid-Eocene (e.g.,
967 Patriat and Achache, 1984, Gibbons et al., 2015, Parsons et al., 2021). Key events during the
968 reorganization (Fig. 10c) included (1) a deceleration of Indian plate motion by 30%–38% between 45-
969 40 Ma (Molnar and Stock, 2009); (2) reduction in spreading rate and clockwise rotation of spreading
970 azimuth between India and Africa between 52-41 Ma (Patriat and Achache, 1984, Cande et al., 2010),
971 followed by southwest jump of the Central India spreading ridge at ~41 Ma (Torsvik et al., 2013); (3)
972 coupling of the Indian and Australian plates following a cessation of spreading on intervening Wharton
973 ridge at ~43-36 Ma (Jacob et al., 2014, Gibbons et al., 2015); (4) onset of northward motion of the
974 Australian plate at ~45-43 Ma, after being stationary since the mid-Cretaceous (Torsvik et al., 2008,
975 Müller et al., 2019); (5) onset of subduction of the Australian plate beneath Java at 47-44 Ma (Smyth

976 et al., 2008); and (6) acceleration of plate spreading between the Australian and Antarctic plates at ~47-
977 45 Ma (Torsvik et al., 2008, Eagles, 2019).

978 Plate reconstructions and mantle tomographic analyses propose that this reorganisation occurred in
979 response to the India-Asia collision (Replumaz et al. 2014; Gibbons et al., 2015, Parsons et al., 2021);
980 however, these studies lacked an unambiguous and precise marker linking the bedrock record of the
981 collision to plate network reorganization. Crucially, our precise age constraint for the onset of UHP
982 exhumation at 47-46 Ma corroborates those previous studies by demonstrating a viable, causative link
983 between changes in the subduction dynamics of the India-Asia collision zone starting at 47-46 Ma (Fig.
984 9) and the resulting mid-Eocene plate network reorganization that immediately followed (Fig. 10c). Our
985 synthesis of the bedrock record of metamorphism and magmatism during the Himalayan orogeny
986 demonstrate that UHP exhumation was rapidly, and in some places immediately, followed by
987 widespread, hotter, amphibolite facies metamorphism (Fig. 9a), which we attribute to shallowing and
988 underthrusting of the Indian plate beneath Eurasia. Consequently, we argue that our new
989 geochronological constraints not only date the onset of UHP exhumation in the NW Himalaya, but also
990 date the changes in subduction dynamics of the collision zone that triggered the Eocene plate network
991 reorganization.

992

993 **6.3. The geodynamic significance of UHP exhumation for orogenesis and plate tectonics**

994

995 Our study demonstrates how our understanding of the bedrock record of continental collisions may be
996 enriched by placing it within the context of plate kinematics. At the same time, the bedrock record
997 provides a crucial means of ground truthing lithospheric- to planetary-scale models of plate tectonics
998 and mantle geodynamics. Further understanding may therefore be gained through new applications of
999 similar integrated approaches to other collision zones.

1000 Multiple examples of HP and UHP metamorphism are recorded across much of SE Asia during the
1001 Early to Late Cretaceous, but their timing and tectonic significance is poorly constrained. This
1002 metamorphism broadly overlaps with multiple terrane accretion events from the region (e.g., Woyla,
1003 Southwest Borneo, and Argoland blocks Hall, 2012, Morley, 2012, Metcalfe, 2021, Advokaat and van
1004 Hinsbergen, 2024, van de Lagemaat and van Hinsbergen, 2024), as well as a previously recognised
1005 plate network reorganization at ~110-100 Ma (Matthews et al., 2012, Müller et al., 2016). Detailed
1006 petrochronology studies of these occurrences of HP/UHP metamorphism are yet to be conducted but
1007 may provide important constraints for plate reconstructions and orogenic models of SE Asia which
1008 continue to be debated and revised (e.g., Hall, 2012, Morley, 2012, Metcalfe, 2021, Advokaat and van
1009 Hinsbergen, 2024, van de Lagemaat and van Hinsbergen, 2024).

1010 Further back in time, there are many examples HP and UHP metamorphic rocks associated with the
1011 Palaeozoic tectonic evolution of the Rheic and Iapetus Oceans and related collisional events including
1012 the Caledonian, Scandian, and Grampian orogenies (see review of Domeier, 2016). The tectonic and
1013 geodynamic significance of these HP/UHP rocks is often unclear. Integrated petrochronology-plate
1014 kinematic analyses of these rocks may better constrain both the tectonometamorphic evolution of these
1015 HP/UHP metamorphic events and the plate kinematic evolution of the surrounding plate network (e.g.,
1016 Gilotti et al., 2008, Hacker et al., 2010, Bottrill et al., 2014, Domeier, 2016). Whilst plate reconstructions
1017 of pre-Jurassic events are less well constrained than those of younger ages, integration of plate
1018 kinematics and bedrock dataset still has potential to reveal new information for older tectonic events
1019 that would remain otherwise hidden when conducting either discipline (bedrock analysis or plate
1020 reconstruction modelling) in isolation.

1021

1022 **7. Summary**

1023 By associating accessory phase ages with distinct metamorphic assemblages, and combining both high
1024 precision and high spatial resolution techniques, we demonstrate that the phase of peak zircon
1025 crystallization recorded in the Tso Morari complex at 47-46 Ma, corresponds to the onset of exhumation
1026 from UHP conditions. This study also describes the record of subduction-related major and accessory
1027 mineral growth preserved in a rare garnet-bearing Puga Gneiss, which records a part of the P-T-t history
1028 not preserved elsewhere. Zircon from a mafic eclogite have a U-Pb CA-ID-TIMS age of 46.912 ± 0.068
1029 Ma (2σ) and an LA-ICPMS age of 47.5 ± 1.7 Ma, with REE profiles indicative of zircon crystallization
1030 at eclogite facies conditions. Those ages overlap with zircon rim ages (48.9 ± 1.2 Ma, LA-ICP-MS) and
1031 xenotime ages (47.4 ± 1.4 Ma; LA-ICP-MS) from the hosting Puga gneiss, which grew during
1032 breakdown of UHP garnet rims, as indicated by garnet element maps. Subsequent exhumation through
1033 the rutile closure temperature to crustal conditions is constrained by new dates of 40.4 ± 1.7 Ma and
1034 36.3 ± 3.8 (2σ LA-ICP-MS).

1035 The overlap between our mafic eclogite zircon ages and our xenotime-UHP garnet break down ages,
1036 indicate that the pulse of zircon growth recorded in the Tso Morari complex at 47-46 Ma (e.g.,
1037 Donaldson et al., 2013) took place as a result of fluid infiltration at the onset of exhumation from UHP
1038 conditions, rather than as a result of peak UHP metamorphism. These ages from Tso Morari overlap
1039 with U-Pb ID-TIMS, SHRIMP, and SIMS analyses of zircon from eclogite-facies mafic rocks in
1040 Kaghan and Naran, ~450-480 km to west of Tso Morari, which yielded ages of 46.4 ± 0.1 Ma, $46.2 \pm$
1041 0.7 Ma, and 46 ± 2 Ma, respectively (Kaneko et al., 2003; Parrish et al., 2006; Zhang et al., 2022). We
1042 interpret this overlap as an indication that exhumation from UHP conditions occurred synchronously at
1043 46-47 Ma across the whole of the NW Himalaya.

1044 Integration of our new ages plus previously published ages from the NW Himalaya with existing
1045 metamorphic, magmatic, and plate kinematic constraints demonstrates that UHP exhumation at 47-46
1046 Ma occurred in response to changes in subduction dynamics following the onset of the India-Asia
1047 collision (*sensu stricto*) between 50-47 Ma. Our data suggest that UHP exhumation was rapidly and in
1048 some places immediately followed by widespread, hotter, amphibolite facies metamorphism which we
1049 attribute to buoyancy-driven shallowing and under thrusting of the continental Indian slab beneath
1050 Eurasia beginning at 47-46 Ma. At a broader perspective, these changes in subduction dynamics and
1051 convergence mechanisms of the Himalayan orogeny (Fig. 9a-b) also impacted the geodynamics and
1052 kinematics of the encompassing tectonic plate network.

1053 Our new data from Tso Morari demonstrably tie the bedrock record of the Himalayan orogeny to slab
1054 dynamics and plate kinematics of the collision zone. Continent-continent collision of India and Asia at
1055 50-47 Ma not only provided the trigger for UHP exhumation, but also resulted in, (1) significant changes
1056 in the metamorphic and magmatic evolution of the Himalayan orogen (Fig. 9a); (2) the transition of the
1057 Himalaya from an accretionary orogen to a collisional orogen (Fig. 9b); and (3) a significant
1058 reorganisation of the wider India-Eurasia-Australia plate network (Fig. 9c). Therefore, our new
1059 constraints not only date the onset of UHP exhumation in the NW Himalaya, but also date changes in
1060 subduction dynamics of the collision zone that triggered the Eocene plate network reorganization. As a
1061 proof-of-concept, our study demonstrates the value of integrated bedrock-plate kinematic approaches,
1062 which ultimately lead to a more holistic and accurate understanding how bedrock geology and plate
1063 tectonics are linked across subgrain to planetary scales.

1064 **8. Supporting Information**

1065 The EPMA, LA-ICPMS and ID-TIMS data used for geochemistry analysis and U-Pb dating in this
1066 study has been uploaded to the Mendeley data repository with a DOI and an open access license CC
1067 BY 4.0. EPMA data DOI: [10.17632/vcfbxxm6h9f.1](https://doi.org/10.17632/vcfbxxm6h9f.1). Geochronology data DOI: [10.17632/t5fd58m67p.1](https://doi.org/10.17632/t5fd58m67p.1).

1068 8.1. Supporting Information 1. Geochronology of the northwest Himalaya

1069 8.1.1. Table A1: Published geochronology results from Tso Morari and Kaghan.

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1081

1082 **10. References**

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