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**Pre-Cenozoic evolution of the Aghil Range (western Tibetan Plateau): A missing piece of the Tibet-Pamir-Karakorum geopuzzle**

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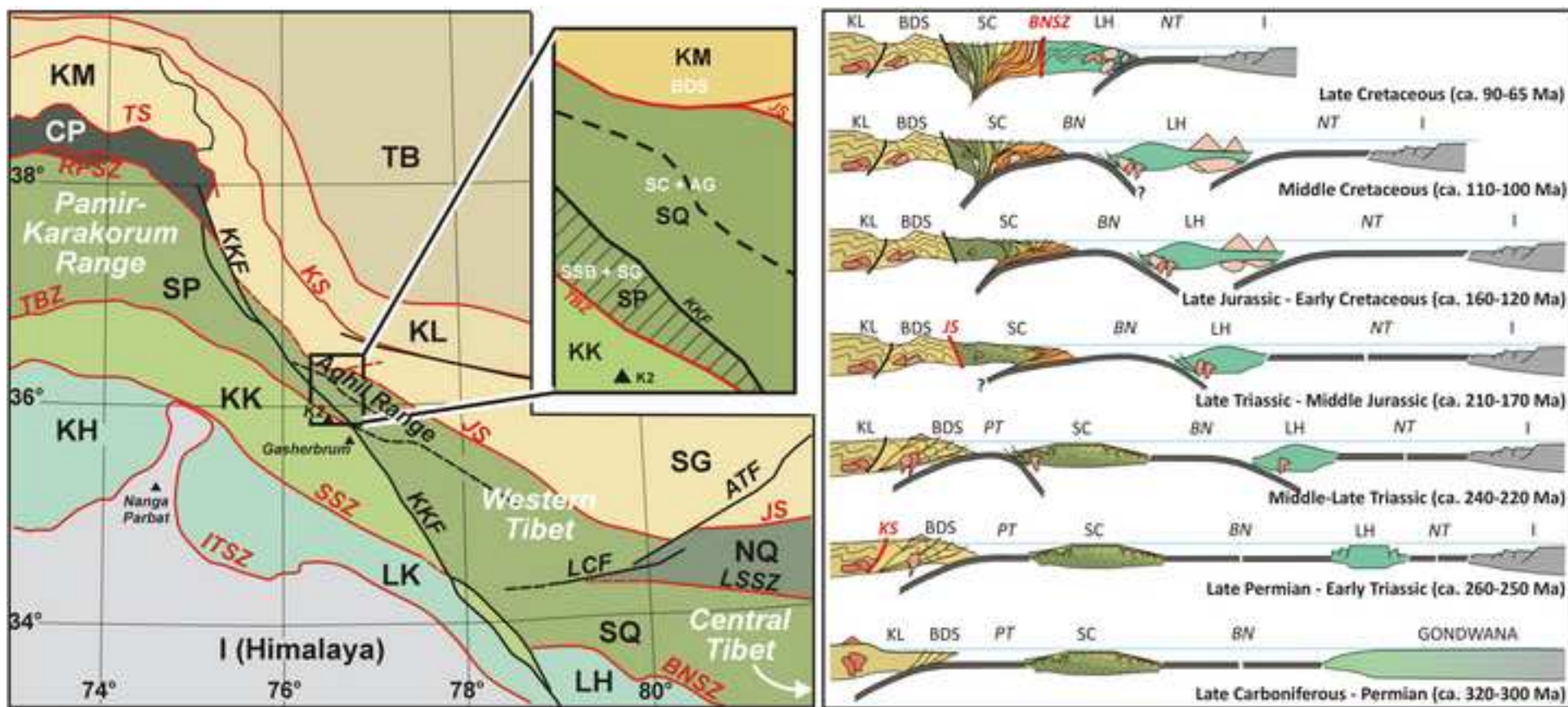
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(Article begins on next page)



Research highlights:

- The pre-Cenozoic evolution of the Aghil Range (western Tibet) is investigated
- A coherent slice of Neoproterozoic basement with its sedimentary cover is preserved
- A correlation between terranes of Central Tibet and Pamir-Karakorum is proposed

1 **Pre-Cenozoic evolution of the Aghil Range (western Tibetan Plateau): a missing piece of**  
2 **the Tibet-Pamir-Karakorum geopuzzle**  
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23 **Abstract**

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24 The Tibetan Plateau, largely derived from the accretion of several Gondwana microplates to the southern  
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4 margin of Asia since the late Palaeozoic, is the highest and largest topographic relief on Earth. Although the  
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6 first order geodynamic processes responsible for its pre-Cenozoic evolution are quite well-known, many  
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8 issues are still debated, among which is the timing of collision of each terrane with the southern margin of  
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10 Asia. Even more uncertain is the pre-Palaeozoic history of these terranes, due to the lack of basement  
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12 exposures. As a contribution to understanding the pre-Cenozoic evolution of the Tibetan Plateau, this paper  
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14 focuses on the Aghil Range, a remote and poorly investigated area close to the Karakorum Fault between  
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16 Kunlun and Karakorum (Xinjiang, China) in western Tibet. The tectono-metamorphic and magmatic  
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18 evolution of the Aghil Range is investigated using a multidisciplinary approach that combines field mapping,  
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20 petrology and geochronology (U-Pb on titanite, zircon, monazite and xenotime using SHRIMP-RG). We  
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22 demonstrate that the Aghil Range preserves a coherent slice of Neoproterozoic crystalline basement with a  
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24 late Palaeozoic sedimentary cover deposited on a passive continental margin during the Gondwana break-  
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26 up. This represents the westernmost exposure of Precambrian crystalline basement known so far in the  
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28 Tibetan Plateau. Furthermore, petrological and geochronological results allow reconstructing the Mesozoic  
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30 poly-metamorphic evolution of this sector of the Tibetan Plateau, which records the evidence of Middle  
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32 Jurassic (ca. 170 Ma) and Late Cretaceous (66 Ma) collisional events, as well as of the Late Jurassic (ca. 150  
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34 Ma) early subduction of an accretionary complex developed on its southern margin. Evidence of Late  
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36 Cretaceous subduction-related magmatism preceding the last collisional event is also recorded. These  
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38 results allow tentative correlation of the different terranes of Central Tibet with those of the Pamir-  
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40 Karakorum Range on both sides of the Karakorum fault.

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41 **Key-words**

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43 Tibetan Plateau, pre-Cenozoic evolution, Cimmerian orogeny, petrology, U-Pb geochronology

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45 **Research highlights**

- 46 • The pre-Cenozoic evolution of the Aghil Range (western Tibet) is investigated
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56 **1. Introduction**

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The Tibetan Plateau, together with the Pamir-Karakorum Range, is the highest and largest topographic plateau on Earth: understanding its formation and evolution is therefore fundamental for clarifying the geodynamic processes leading to crustal thickening and continental growth. The origin, evolution and tectonic architecture of the Tibetan Plateau and of the Pamir-Karakorum Range have been studied for several decades (e.g. Sengör, 1979, 1987; Allégre et al., 1984; Sengör, 1987; Dewey et al., 1988; Gaetani et al., 1990a,b, 1993; Burtman and Molnar, 1993; Matte et al., 1996; Gaetani, 1997; Yin and Harrison, 2000); important advances in understanding their genesis have been made over the last ten years thanks to an increasing number of field, geochemical and geochronological studies as demonstrated by publication of numerous review papers (e.g. Zhang et al., 2012; Zhu et al., 2011, 2013) and special volumes dedicated to this topic (e.g. Zhang and Santosh, 2012; Zanchi et al., 2015; Chung [and Niu et al.](#), 2016; Zhang et al., 2017).

It is now widely accepted that the Tibetan Plateau and Pamir-Karakorum Range derive from the accretion of several Gondwana-derived microplates (also called Cimmerian terranes: Sengör, 1984) to the southern margin of Asia since the late Palaeozoic, in response to the ongoing subduction and progressive closure of the oceanic basins between each terrane (e.g. Allégre et al., 1984; Dewey et al., 1988; Yin and Harrison, 2000; Pan et al., 2012). Moreover, there is increasing evidence that most of the deformation, shortening and crustal thickening within the plateau were the result of these pre-Cenozoic accretionary processes whereas the contribution of India-Asia collision to the building of the plateau was only minor (e.g. Murphy et al., 1997; Yin and Harrison, 2000; Hildebrand et al., 2001; Robinson et al., 2004; Kapp et al., 2003b, 2005, 2007; Guynn et al., 2006; Zhang et al., 2012; Zhu et al., 2013). However, several recent studies explain crustal thickening as related to the underthrusting of Indian lithosphere (e.g. Chen et al., 2017). Although the first order geodynamic processes responsible for the pre-Cenozoic evolution of the plateau are quite well-known, many issues are still debated. These include the timing, duration and direction of oceanic subduction and the timing of collision of each terrane with the southern margin of Asia. Even more uncertain is the Precambrian history of the Cimmerian terranes, due in large part to the scarcity of basement exposures and the predominance of late Palaeozoic or younger supracrustal assemblages (e.g. Pan et al., 2004; Guynn et al., 2012; Zhu et al., 2013 and references therein).

Most studies devoted to understanding the pre-Cenozoic formation and evolution of the Tibetan Plateau and Pamir-Karakorum Range are based on either stratigraphic evidence (e.g. Gaetani et al., 1990b, 1993; Gaetani, 1997; Kapp et al., 2007; Zanchi and Gaetani, 2011; Zanchi et al., 2012; Angiolini et al., 2013, 2015; Gaetani and Leven, 2014; Robinson, 2015; Zeng et al., 2016) or on the distribution, composition and age of subduction- and/or collision- related magmatic rocks (e.g. Schwab et al., 2004; Zhu et al., 2011, 2013; Zhang et al., 2012 and references therein; Zanchetta et al., 2018). Studies of metamorphic rocks are sporadic and limited to few areas, such as the Central Qiangtang Metamorphic Belt (e.g. Kapp et al., 2000,

2003a; Pullen et al., 2008, 2011; Liang et al., 2012; Zhai et al., 2011b; Zhao et al., 2014), the Amdo terrane (e.g. Guynn et al., 2006, 2012, 2013; [Zhang et al., 2012a](#)), the Central Lhasa terrane (e.g. Dong et al., 2011; [Zhang et al., 2012b](#); [Zhang et al., 2014](#)) and the southern Karakorum Terrane (e.g. Searle and Tirrull, 1991; Lemennicier et al., 1996; Fraser et al., 2001; Streule et al., 2009; Searle, 2011).

As a contribution to the understanding of the pre-Cenozoic evolution of the Tibetan Plateau, this study focuses on the metamorphic and magmatic units exposed along the Aghil Range, a remote and poorly investigated area located at the junction between the Tibetan Plateau and the Pamir-Karakorum Range in western Tibet. Interpretation of this area is still controversial, having been alternatively ascribed to the Songpan-Ganze (Tianshuai) terrane (e.g. Valli et al., 2008; Leloup et al., 2012; Pan et al., 2012) or to the Southern (or Western) Qiangtang terrane (e.g. Gaetani et al., 1990a; Robinson, 2009, 2015; Groppo and Rolfo, 2008; Streule et al., 2009; Searle et al., 2010, 2011; Gaetani and Leven, 2014; Rolfo et al., 2014). The aim of this study is therefore twofold: (i) to understand the geological significance of the Aghil Range and assess the nature of the boundaries between the various terranes on both sides of the Karakorum Fault; (ii) to tentatively reconstruct the pre-Cenozoic history of accretion, collision, metamorphism and magmatism of the western portion of the Tibetan Plateau, in comparison with the better known evolution of the central portion of the plateau.

We present field, petrographic, petrologic and geochronologic data on both metamorphic and magmatic rocks exposed along a geological transect, approximately 40 km long, located between the Kunlun to the north and the Karakorum to the south (Xinjiang, China). The studied metamorphic rocks belong to two different tectonic units (Gaetani et al., 1990a): the “Bazar Dara Slates”, a metasedimentary unit located few km south of Mazar (not to be confused with the thick terrigenous Bazardara Series of SE Pamir, which are tentatively correlated with the Singhiè Formation of the Shaksgam Sedimentary Belt by Gaetani and Leven, 2014), and the “Surukwat Complex”, a composite sequence of metamorphic thrust sheets derived from both magmatic and sedimentary protoliths. Two granodioritic bodies (i.e. Aghil Granodiorite and Sughet Granodiorite) tectonically interposed between these metamorphic units have been investigated as well.

The petrological and geochronological results allow us to: (i) clarify how the Cimmerian terranes are assembled on both sides of the Karakorum fault (i.e. in the western Tibetan Plateau and in the Pamir-Karakorum Range); (ii) demonstrate that the Surukwat Complex represents the westernmost exposure of a Precambrian crystalline basement known so far in the Tibetan Plateau; (iii) reconstruct the Mesozoic poly-metamorphic evolution of both the Bazar Dara Slates Unit and the Surukwat Complex which record the evidence of Middle Jurassic and Late Cretaceous collisional events, as well as the Late Jurassic early subduction of an accretionary complex developed on the southern margin of the Surukwat Complex; (iv) recognise evidence of Late Cretaceous subduction-related magmatism preceding the collision of the Lhasa and South Qiangtang terranes; and (v) confirm the diachronicity, from east to west, of the Cretaceous

126 collisional event between the Lhasa and South Qiangtang terranes in western Tibet and between the  
127 Kohistan-Ladakh and Karakorum terranes in Pamir-Karakorum .

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## 130 **2. Geodynamic Setting of the Tibetan Plateau and Pamir-Karakorum Range**

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### 133 **2.1 Tibetan Plateau**

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136 The Tibetan Plateau consists of four main E-W trending crustal terranes that rifted from the eastern margin  
137 of Gondwana during the late Palaeozoic, drifted northward across the Tethyan Ocean basins and then  
138 progressively accreted to the southern margin of Asia during the Mesozoic (e.g. Allégre et al., 1984; Sengör,  
139 1987; Yin and Harrison, 2000). In central Tibet, the following terranes are conventionally distinguished from  
140 north to south (Fig. 1a): Kunlun, Songpan-Ganze, Qiangtang and Lhasa terranes. The boundaries between  
141 these terranes coincide with different suture zones resulting from the closure of the ocean basins originally  
142 interposed between each terrane, and now marked by discontinuous belts of ophiolite fragments and  
143 m $\acute{e}$ lange:

- 144 – The Jinsha Suture (JS) separates the Songpan-Ganze terrane from the Qiangtang terrane and records  
145 closure of the Paleo-Tethys Ocean; the Songpan-Ganze terrane is commonly interpreted as an extensive  
146 arc-accretionary system built along the southern margin of Kunlun during the Triassic (e.g. Matte et al.,  
147 1996; Schwab et al., 2004), whose huge volume of sediments did not allow a complete continent-  
148 continent collision between Kunlun and Qiangtang (e.g. Roger et al., 2010).
- 149 – The Bangong-Nujiang Suture Zone (BNSZ), more than 1200 km long, separates the Qiangtang terrane  
150 from the Lhasa terrane and resulted from the closure of the Bangong-Nujiang Ocean (or Meso-Tethys  
151 Ocean; Sengör, 1984). In western and eastern Tibet, the BNSZ ophiolitic belt is doubled and isolates two  
152 micro-blocks interposed between Qiangtang and Lhasa terranes: the Risum block to the west and the  
153 Amdo terrane to the east. The Risum block is interpreted as an oceanic arc formed by the intra-oceanic  
154 subduction of the Bangong-Nujiang Ocean (Matte et al., 1996; Shi et al., 2004; Shi, 2007; Liu D. et al.,  
155 2017). The Amdo terrane is an old micro-continent within the Bangong-Nujiang Ocean that  
156 amalgamated with the Qiangtang block before the Lhasa–Qiangtang collision (Xu et al., 1985; Guynn et  
157 al., 2006). The Amdo block has been variably correlated to the Lhasa terrane (e.g. Coward et al., 1988;  
158 Harris et al., 1988; Yin and Harrison, 2000; Pan et al., 2004) or the Qiangtang terrane (e.g. Guynn et al.,  
159 2006, 2012; Zhu et al. 2013).
- 160 – The 2000-km-long Indus-Tsangpo Suture Zone (ITSZ) represents the site where the Neo-Tethys  
161 lithosphere separating the Lhasa terrane and north India was consumed at a subduction zone dipping  
162 northward beneath the Lhasa terrane (Yin and Harrison, 2000).

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160 – The Longmu Tso-Shuanghu Suture Zone (LSSZ) divides the Qiangtang terrane in two sub-terrane: the  
161 North (or Eastern) Qiangtang terrane and the South (or Western) Qiangtang terrane. This suture zone is  
162 spatially associated with a high-pressure metamorphic belt (Central Qiangtang Metamorphic Belt,  
163 CQMB: Kapp et al., 2000, 2003a), and its origin is still debated. The CQMB has been interpreted either:  
164 (i) as a part of the Songpan-Ganze accretionary mélange that was underthrust beneath a single  
165 Qiangtang terrane during the southward subduction of the Paleo-Tethys along the Jinsha Suture in the  
166 early Mesozoic (i.e. underthrust model: e.g. Kapp et al., 2000, 2003a; Kapp, 2001; Pullen et al., 2011), or  
167 (ii) as an *in situ* suture zone formed by northward subduction of a branch of Paleo-Tethys originally  
168 separating the South Qiangtang terrane of Gondwanan affinity from the North Qiangtang terrane of  
169 Cathaysian affinity (i.e. intra-Qiangtang suture model: e.g. Zhang, 2001; Zhang et al., 2006a,b, 2011;  
170 Zhang and Tang, 2009; Liu et al., 2011; Zhai et al., 2011a,b, 2013; Zhu et al., 2013; Zhao et al., 2014).

171 The timing, duration and direction of oceanic subduction, as well as the timing of final collision between  
172 each terrane are still debated. The Paleo-Tethys ocean was subducted northward beneath the Kunlun and  
173 Songpan-Ganze terranes by the Late Permian - Early Triassic, as evidenced by the occurrence of a magmatic  
174 belt and volcanic arc along the southern margin of Kunlun (e.g. Matte et al., 1996; Xiao et al., 2002, 2003;  
175 Liu et al., 2015; Cao et al., 2015); southward subduction in Late Triassic – Early Jurassic times is instead  
176 proposed by other studies (e.g. Kapp et al., 2000, 2003a; Zhang et al., 2016). The final collision between  
177 the Qiangtang and the Songpan-Ganze/Kunlun terranes occurred in the Late Triassic to Middle Jurassic. The  
178 southernmost Bangong-Nunjiang ocean was subducted either northward under the South Qiangtang  
179 terrane, or southward beneath the Lhasa Terrane during the Mesozoic (Zhu et al., 2013; Liu D. et al., 2017  
180 and references therein). This was likely an oblique subduction, resulting in a diachronous collision of the  
181 Lhasa and Qiangtang terranes from west (Middle Cretaceous) to east (Late Cretaceous) (e.g. Matte et al.,  
182 2006; Zhang et al., 2008; Zhu et al., 2013; Liu L. et al. 2017; Liu D. et al., 2017 and references therein).

## 184 **2.2 Pamir-Karakorum Range**

185 The dextral strike-slip Karakorum fault (KKF) (e.g. Phillips et al., 2004; Searle and Phillips, 2007; Phillips,  
186 2008; Valli et al., 2008; Robinson, 2009; Leloup et al., 2011) separates the Tibetan Plateau to the east from  
187 the Pamir-Karakorum Range to the west, whose tectonic framework is also the result of the amalgamation  
188 of different Gondwana-derived terranes. From north to south these terranes are the Kunlun, Karakul-Mazar  
189 (or North Pamir), Central Pamir, South Pamir, Karakoram and Kohistan-Ladakh, and they are bounded by  
190 the Kunlun, Tanyamas, Rushan-Pshart, Tirich Boundary Zone and Shyok suture zones, respectively (Fig. 1a).  
191 The correlation between crustal terranes and suture zones of Pamir-Karakorum and central Tibet is not  
192 univocal (e.g. Burtman and Molnar, 1993; Yin and Harrison, 2000; Lacassin et al., 2004; Schwab et al., 2004;  
193 Robinson, 2009, 2015; Robinson et al., 2012; Zanchetta et al., 2018). The Karakul-Mazar terrane, bounded

195 by the Kunlun suture to the north and the Tanymas Suture to the south, is commonly considered the  
196 equivalent of the Songpan-Ganze terrane of central Tibet. Interpretation of the other terranes is less  
197 certain. Recent studies suggest that the Central Pamir, South Pamir and Karakoram terranes are correlative  
198 to the Qiangtang terrane (e.g. Phillips et al., 2004; Upadhyay et al., 2005; Searle and Phillips, 2007;  
199 Robinson et al., 2004; Robinson, 2009, 2015; Searle et al., 2010; Searle, 2011; Angiolini et al., 2013; Yang et  
200 al., 2017) and that the Kohistan-Ladakh terrane and the Lhasa terrane of central Tibet are part of a  
201 continuous magmatic arc, built on a continental basement in the Lhasa terrane and in the Ladakh region,  
202 and on oceanic crust in the Kohistan terrane (e.g. Rolfo et al., 1997; Yin and Harrison, 2000; Robinson et al.,  
203 2004; Robinson, 2009, 2015). Following this interpretation, the Shyok Suture, that separates the Karakorum  
204 terrane from the Kohistan-Ladakh terrane, would be the equivalent of the Bangong-Nunjiang suture zone.  
205 However, other studies suggest that South Pamir is the equivalent of the Lhasa terrane (e.g. Lacassin et al.,  
206 2004; Schwab et al., 2004), in which case the Rushan-Pshart Suture Zone would be equivalent to the  
207 Bangong-Nunjiang suture zone. An absence of direct correlations between major suture zones east and  
208 west of the Karakorum Fault has been also recently proposed (Zanchetta et al., 2018). A better  
209 understanding of the western Tibet region and the Aghil Range where the study area is located in  
210 particular, is crucial for meaningful correlation of crustal terranes across the Karakorum fault.

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### 3. Geological Setting

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215 The Aghil Range in western Tibet (Fig. 1a) was studied along a cross-section between Ilik (at the confluence  
216 between the Yarkand and the Zug Shaksgam rivers) and Sughet Jagal (northern K2 base camp, along the  
217 Sarpo Laggo Valley), across the Aghil pass and the Shaksgam Valley (Fig. 1b). Very few geological studies  
218 have been performed in this area, tectonically sandwiched between Kunlun to the north and Karakorum to  
219 the south. The first “modern”, though preliminary, data are reported in Gaetani et al. (1990a, 1991); these  
220 results already documented the complexity of the area, characterized by the juxtaposition of different  
221 metamorphic, sedimentary or magmatic units. More recently, Groppo and Rolfo (2008) reported the  
222 evidence of a possibly old metamorphic basement north of Aghil Pass (i.e. the Surukwat Complex), and  
223 constrained the P-T evolution of its structurally upper portion. However, the nature and age of this  
224 basement is still unknown, as well as the age of metamorphism. The non-metamorphic Shaksgam  
225 Sedimentary Belt was investigated in detail by Gaetani et al. (1990a, 1991) and Gaetani and Leven (2014),  
226 whereas the magmatic rocks exposed close to the Aghil Pass (i.e. the Aghil Granodiorite) were studied very  
227 recently by Liu L. et al. (2017). Detailed description of the lithological and tectonic architecture along the Ilik  
228 – Sughet Jagal transect and relevant images of field geology and mesoscopic structures are given by Rolfo  
229 et al. (2014) and will only be summarized here.

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229 Starting from Ilik and going upstream (southward) along the Aghil Dara Valley, the Bazar Dara Slates  
230 Unit (Fig. 1b) is a metasedimentary sequence consisting mainly of phyllites and metasandstones that are  
231 dipping steeply towards SSE and locally intruded by undeformed Late Triassic granitic bodies (e.g. Mazar  
232 Granodiorite; Liu et al., 2015). A sub-vertical to south-dipping fault separates the Bazar Dara Slates Unit  
233 from the Surukwat Complex; this is a composite sequence of thrust sheets steeply dipping SSW (Fig. 1b,d)  
234 in which, except for a few non-metamorphic slivers, there is a general southward increase in metamorphic  
235 grade from lower to higher structural levels (Rolfo et al., 2014). Although pervasively mylonitized, the  
236 internal coherence of this basement is mostly preserved (Fig. 1c). From north to south, the Surukwat  
237 Complex starts with a non-metamorphic sliver of red sandstone with anhydrite interlayers that shows a  
238 strong affinity with the Qiangtang terrane (Leeder et al., 1988), and is petrographically similar to the  
239 Jurassic Marpo Sandstone of the Shaksgam Valley (Gaetani et al., 1990a, 1991). A metamorphic basement  
240 derived from igneous protoliths is thrust over the red sandstone to the south (Fig. 1d). From lower to  
241 higher structural levels it consists of meta-diorite, meta-granodiorite and meta-granite with transposed  
242 meta-aplitic dykes. Primary intrusive relationships between the various magmatic protoliths have been  
243 obliterated by the pervasive mylonitic deformation which affected this portion of the Surukwat Complex. A  
244 small (few-meters-thick) slice of slightly metamorphosed limestone is tectonically intercalated in this meta-  
245 igneous basement. A few-kilometers-thick layer of metaconglomerate dominated by granitic and dioritic  
246 clasts follows further upstream. Further south, the metaconglomerate is overlain by a km-thick  
247 metasedimentary sequence that consists of quartzite, quartzitic gneiss and metapelite with meter-thick  
248 intercalations of meta-marl (amphibole-bearing calcschist, biotitic-amphibolitic schist, carbonate-rich  
249 garnet-bearing biotitic schist) and impure marble. The medium grade metapelites represent the last unit of  
250 the Surukwat Complex, which is bounded to the south by a sharp fault contact with the Aghil intrusive body  
251 (Fig. 1b,d). A weakly deformed apophysis of this pluton crops out within the quartzitic gneisses to the  
252 north. The Aghil intrusive body mainly consists of amphibole + biotite –bearing granodiorite, with minor  
253 monzonite and porphyritic granite, the latter cropping out in the proximity of the Aghil Pass. A Late  
254 Cretaceous age has been recently obtained for these magmatic rocks (zircon U-Pb ages; Liu L. et al., 2017).

255 Another tectonic contact is crossed southward just before the Aghil Pass (Fig. 1b,d) that separates  
256 the Aghil Granodiorite from the Shaksgam Sedimentary Belt (Gaetani et al., 1990a, 1991), a ~250 km long  
257 and 15-20 km wide sedimentary sequence exposed along the Shaksgam Valley and extending to the west of  
258 the Karakorum fault for ~150 km in southeast Pamir (Robinson, 2009). This sedimentary sequence,  
259 displaced in a system of open folds, faulted, thrust and stacked together, is at least 3 km thick and spans  
260 from the Lower Permian to Jurassic (Gaetani et al., 1990a, 1991; Gaetani, 1997; Gaetani and Leven, 2014).  
261 The main strand of the Karakorum Fault is crossed a few kilometres south of the junction between the  
262 Sarpo Laggo and Shaksgam valleys (Fig. 1b,d). A cataclastic contact separates the Shaksgam Sedimentary  
263 Belt (Aghil Dolomite) from the Sughet Granodiorite. This plutonic body, mostly made of biotite-bearing

264 granodiorite, crops out in the lower (northern) part of the Sarpo Laggo valley, near Sughet Jagal, and  
265 represents the last unit of the investigated geological transect.

266 This study focuses on six representative samples from the Bazar Dara Slates Unit (BDS: sample 06-  
267 01), the Surukwat Complex (SC: samples 06-10, 06-17 and 06-115) and the Aghil and Sughet magmatic  
268 bodies (samples 06-26 and 06-108). These samples have been selected out of a total of 76 samples (BDS; 16  
269 samples; SC: 54 samples; magmatic bodies: 6 samples) after careful petrographic characterization of the  
270 entire sample suite.

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## 4. Methods

### 4.1 Mineral chemistry

Minerals were analysed with a Cambridge Stereoscan 360 SEM equipped with an EDS Energy 200 and a Pentafet detector (Oxford Instruments) at the Department of Earth Sciences, University of Torino. The operating conditions were as follows: accelerating voltage was set to 15 kV, beam diameter was 2  $\mu\text{m}$ , and detection limits for oxides were 0.03 wt%. SEM–EDS quantitative data were acquired and processed using the Microanalysis Suite Issue 12, INCA Suite version 4.01; natural mineral standards were used to calibrate the raw data; the  $\rho\phi Z$  correction (Pouchou and Pichoir, 1988) was applied. Absolute error is 1  $\sigma$  for all calculated oxides. Mineral chemical data of representative minerals are reported in Table 1.

### 4.2 Determination of peak P-T conditions in the metamorphic samples

P-T conditions for all the metamorphic samples were estimated using the “Average PT” routine of THERMOCALC (Holland and Powell, 1998, version 3.33, thermodynamic database 5.5). Activity-composition relationships were calculated using AX. This method, which estimates the optimal P-T conditions using a set of independent reactions that fully describe the thermodynamics of the system, is able to find a result only if the given mineral assemblage defines a sufficient number of reactions between end-members. The obtained results were considered reliable if passed the ‘sigfit’ test ( $\sigma_{\text{fit}} < \text{cutoff value}$ ), giving P–T uncertainties ( $\sigma T$  and  $\sigma P$ ) at  $\pm 1\sigma$  (95% confidence). End member(s) with erratic behaviour (large  $e^*$  values) and a low influence on the least squares results (low *hat* values) were removed from the calculation because they may cause inconsistency in the results (see Powell and Holland, 1994). A pure H<sub>2</sub>O fluid was considered in the calculations. Average P-T results are reported in Table 2.

The pseudosection approach cannot be applied on most of the studied samples. Sample 06-1 (Bazar Dara Slates phyllite) contains significant amounts of calcite, mostly concentrated in late veins, and it is not

299 possible to obtain the equilibrium bulk composition needed for the pseudosection calculation. Similarly, an  
300 equilibrium bulk composition cannot be determined for samples 06-10 and 06-17 (meta-granodiorite and  
301 meta-diorite from the Surukwat Complex), which are clearly not completely re-equilibrated, as suggested  
302 by the presence of relict magmatic hornblende crystals (see section 4.1.2). Therefore, the pseudosection  
303 approach has been applied only on sample 06-115 (two-micas, garnet-bearing micaschist from the  
304 Surukwat Complex). The pseudosection for this sample was already calculated by Groppo and Rolfo (2008)  
305 and it was used to constrain its complete P-T evolution.

### 307 **4.3 Geochronology and Trace Element Analyses**

309 U/Pb geochronology was performed on zircon separated from igneous samples (06-26, 06-108) and  
310 titanite, monazite, and xenotime in-situ on polished thin sections of metamorphic rocks (06-1, 06-10, 06-17,  
311 06-115), using the SHRIMP-RG (sensitive high-resolution ion microprobe with reverse geometry) instrument  
312 at the Stanford–U.S. Geological Survey Micro-Analysis Center at Stanford University.

313 Heavy mineral separates from granitoid samples 06-26 and 06-108 were obtained by standard  
314 pulverizing, magnetic and heavy liquid methods. Individual zircon grains were handpicked under alcohol,  
315 mounted in epoxy resin with natural zircon standards and polished to expose the grain centers for analysis  
316 by secondary ion microprobe spectrometry (SIMS). Zircon grains were imaged by cathodoluminescence (CL)  
317 to expose intra-grain zoning or complexity and aid in placing SIMS spots. The U-Pb and trace element  
318 analysis (Tables 3 and SM1a) was performed simultaneously following routines outlined in Barth and  
319 Wooden (2006, 2010). Instrument mass fractionation corrections were calibrated by replicate analysis of  
320 the zircon standard R33 zircon (419 Ma; Black et al., 2004) with a  $2\sigma$  calibration error for the R33  $^{206}\text{Pb}/^{238}\text{U}$   
321 ratio of 0.69% for the analytical session added in quadrature. Ages were calculated from  $^{206}\text{Pb}/^{238}\text{U}$  ratios  
322 corrected for common Pb using the  $^{207}\text{Pb}$  method using measured  $^{207}\text{Pb}/^{206}\text{Pb}$  ratios or using the  $^{204}\text{Pb}$   
323 method (see Williams, 1998). The U concentration was calibrated with Madagascar Green (MAD-559; 3940  
324 ppm U, Coble et al., 2018). Data reduction and plotting utilized programs Squid 2.51 and Isoplot 3.76 of  
325 Ludwig (2009, 2012).

326 The acquisition routine included  $^{89}\text{Y}$ ,  $^{139}\text{La}$ ,  $^{140}\text{Ce}$ ,  $^{146}\text{Nd}$ ,  $^{147}\text{Sm}$ ,  $^{153}\text{Eu}$ ,  $^{155}\text{Gd}$ ,  $^{163}\text{Dy}^{16}\text{O}$ ,  $^{166}\text{Er}^{16}\text{O}$ ,  
327  $^{172}\text{Yb}^{16}\text{O}$ ,  $^{90}\text{Zr}_2^{16}\text{O}$ , and  $^{180}\text{Hf}^{16}\text{O}$  simultaneous with U/Pb analysis. In a separate analytical session, additional  
328 trace element analyses were performed, including  $^{27}\text{Al}$ ,  $^{30}\text{Si}$ ,  $^{31}\text{P}$ ,  $^{39}\text{K}$ ,  $^{40}\text{Ca}$ ,  $^{28}\text{Si}^{16}\text{O}$ ,  $^{45}\text{Sc}$ ,  $^{48}\text{Ti}$ ,  $^{49}\text{Ti}$ ,  $^{56}\text{Fe}$ ,  $^{89}\text{Y}$ ,  
329  $^{93}\text{Nb}$ ,  $^{94}\text{Zr}^{16}\text{O}$ , and  $^{96}\text{Zr}$ . Each isotope was normalized to  $^{28}\text{Si}^{16}\text{O}$  or  $^{90}\text{Zr}_2^{16}\text{O}$ , and concentrations were  
330 calibrated against zircon standard MAD-559 (Coble et al., 2018). The estimated errors based on repeated  
331 analysis of MAD-559 was 6 to 10 % for P, Sc, Ti, Y and Nb. The uncertainty of Al, K, Ca, and Fe were higher  
332 (up to 45% RSD), but these elements were measured only to evaluate if the analytical spot intersected an  
333 inclusion or alteration. For example, grain 0608-2.1, Sc, Nb, and Ti were omitted because Al, K and Fe were

334 ~20 to 100x higher than other zircon from the same sample. Chondrite normalized plots were calculated  
335 using values from McDonough and Sun (1995). The  $^{49}\text{Ti}$  data were used to determine the Ti content to  
336 avoid interference of  $^{96}\text{Zr}^{2+}$  with the  $^{48}\text{Ti}$  peak (Watson and Harrison, 2005). Ti-in-zircon temperatures were  
337 calculated using Ferry and Watson (2007), assuming the activity of  $\text{SiO}_2$  is equal to one ( $a_{\text{SiO}_2} = 1$ ) and  
338 activity of  $\text{TiO}_2$  is approximately 0.7 ( $a_{\text{TiO}_2} = 0.7$ ) for rutile-absent siliceous melts (Hayden and Watson,  
339 2007).

340 Titanite (samples 06-10 and 06-17), monazite (06-115, 06-1), and xenotime (06-115, 06-1) were  
341 analyzed *in-situ* in polished thin sections that were cut into fragments and mounted with appropriate  
342 natural standards in large format epoxy mounts (megamounts). Elemental maps showing U, Y, Ce, P, Ca,  
343 and Th concentrations of monazite and high contrast backscatter electron (BSE) images of titanite,  
344 monazite and xenotime were generated to identify zoning prior to analysis. Monazite element maps were  
345 collected on a JEOL JXA-0823 Electron Microprobe at the University of Iowa and BSE imaging was  
346 performed at Stanford University using a JEOL 5600 SEM.

347 U-Pb analysis of titanite (Tables 4 and SM1b) monazite (Tables 5 and SM1c), and xenotime (Tables 6  
348 and SM1d) followed the same analytical routine used for zircon, except  $^{89}\text{Y}$  was not included in the  
349 acquisition table for xenotime and only U and Th were measured as trace elements for titanite. U-Pb ages  
350 were standardized relative to 44069 monazite (425 Ma; Aleinikoff et al., 2006), MG-1 xenotime (490 Ma;  
351 Fletcher et al., 2010; Aleinikoff et al., 2012), and MMs titanite (524 Ma; Schoene and Bowring, 2006)  
352 reference materials for monazite, xenotime, and titanite unknown samples, respectively. For trace  
353 elements, each isotope was normalized  $^{140}\text{Ce}^{31}\text{P}^{16}\text{O}_2$ ,  $^{89}\text{Y}^{16}\text{O}_2$ , or  $^{40}\text{Ca}^{48}\text{Ca}^{48}\text{Ti}^{16}\text{O}_2$ , and trace element  
354 concentrations were standardized relative to 44069 monazite (calibrated relative to Namibia (NAM)  
355 monazite; Aleinikoff et al., 2012), MG-1 xenotime, or BLR titanite (Aleinikoff et al., 2007) for monazite,  
356 xenotime, and titanite, respectively.

357 Common Pb composition for titanite samples was determined by linear regression of all analyses on  
358 a 3D Tera-Wasserburg plot which yielded a data-defined  $^{207}\text{Pb}/^{206}\text{Pb}$  upper intercept of 0.9096, interpreted  
359 as the best estimate of the common Pb composition. All other mineral use initial common Pb isotopic  
360 composition approximated from Stacey and Kramers (1975). Data reduction for geochronology follows the  
361 methods described by Williams (1998) and Ireland and Williams (2003), using the MS Excel add-in programs  
362 Squid2.51 and Isoplot3.76 of Ken Ludwig (2009, 2012). For titanite, the  $^{206}\text{U}/^{238}\text{U}$  calibration constant  
363 utilized a data-defined slope of 1.24 through the distribution of MMs analyses on a plot of  $\ln(\text{UO}^+/\text{U}^+)$  vs  
364  $\ln(\text{Pb}^+/\text{U}^+)$ . Zircon was calculated with a fixed slope of 2.0. Monazite and xenotime analysis used energy  
365 filtering to eliminate the isobaric interference at mass  $^{204}\text{Pb}$ , and a calibration of  $\ln(\text{UO}_2^+/\text{UO}^+)$  vs  
366  $\ln(\text{Pb}^+/\text{UO}^+)$  with a data-defined slope (0.60 and 2.14, respectively), following methods modified from  
367 Fletcher et al. (2010) and Cross and Williams (2018).

## 369 5. Results

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### 371 5.1 Petrography and petrology of the studied samples

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#### 373 5.1.1 Bazar Dara Slates Unit (sample 06-1)

374 The Bazar Dara Slates Unit exposed along the lower Aghil Dara Valley consists of a metasedimentary  
375 sequence of sandstones, siltstones and slates, steeply dipping towards SSE and locally rich in deformed  
376 quartz + carbonate veins. Sample 06-1, collected ca. 4 km south-east of Ilik (N36°22'28.4" E76°40'54.4" –  
377 3560 m), is a two-micas + chlorite phyllite (meta-sandstone) consisting of quartz, muscovite, biotite,  
378 chlorite, calcite, minor albite and accessory ilmenite (Fig. 2a). Most calcite is a late phase, concentrated in  
379 millimetric veins either concordant or discordant with respect to the main foliation; however, it cannot be  
380 excluded that minor calcite is also present in the equilibrium assemblage. The main foliation ( $S_m$ ), defined  
381 by the preferred orientation of muscovite ( $Si = 3.13-3.25$  a.p.f.u. on the basis of 11 oxygens), chlorite ( $X_{Mg} =$   
382  $0.53-0.54$ ), biotite ( $X_{Mg} = 0.51-0.54$ ;  $Ti = 0.11-0.12$  a.p.f.u.) and ilmenite, is pervasively crenulated, with the  
383 local appearance of an  $S_{m+1}$  defined by white mica and ilmenite. Minor monazite and xenotime occur as  
384 anhedral grains with no clear relationship relative to the dominant foliation (Fig. 3).

385 Equilibrium P-T conditions for this sample were calculated for the  $S_m$  assemblage  $Qz + Ab + Chl +$   
386  $Mu + Bt + Ilm$ , which resulted in  $320 \pm 32$  °C,  $5.2 \pm 0.9$  kbar (i.e. greenschist-facies conditions) (Table 2). The  
387 rare occurrence of relict phengite flakes ( $Si = 3.32-3.39$  a.p.f.u.) partially replaced by muscovite suggests a  
388 complex metamorphic evolution, possibly characterized by an earlier high-pressure stage (pre- $S_m$ :  $Qz + Ab +$   
389  $Chl + Phe$ ).

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#### 391 5.1.2 Surukwat Complex (samples 06-10, 06-17, 06-115)

##### 392 4.1.2.1 Meta-diorite and meta-granodiorite (samples 06-10 and 06-17)

393 The lowermost portion of the Surukwat Complex consists of a sequence of strongly mylonitized  
394 metabasites of dioritic to granodioritic composition with sub-vertical attitude, alternating with granitic to  
395 aplitic layers (Fig. 1d). Sample 06-10 and 06-17 are representative examples of the most and less deformed  
396 lithologies, respectively.

397 Sample 06-10 (N36°17'54,8" E76°35'22,8" – 3830 m) is a mylonitized hornblende-bearing meta-  
398 granodiorite still preserving mineralogical relics of the igneous protolith (Fig. 2b). It is characterized by mm-  
399 sized, sharply zoned amphibole porphyroclasts (Fig. 2b,c), with a yellow-pale green relic core (i.e.  $Amp_1$ :  
400 magmatic Mg-hornblende to edenite;  $Si = 6.6-7.1$  a.p.f.u.,  $Al^{IV} = 1.1-1.4$  a.p.f.u.,  $Al^{VI} = 0.2-0.5$  a.p.f.u.,  $X_{Mg} =$   
401  $0.55-0.70$ ) surrounded by a light green rim ( $Amp_2$ : metamorphic actinolite;  $Si = 7.5-7.8$ ,  $Al^{IV} = 0.2-0.6$   
402 a.p.f.u.,  $Al^{VI} = 0.0-0.15$  a.p.f.u.,  $X_{Mg} = 0.65-0.80$ ). A very thin outermost rim of Mg-hornblende ( $Amp_3$ :  $Si =$   
403  $7.2-7.4$  a.p.f.u.,  $Al^{IV} = 0.6-0.8$  a.p.f.u.,  $Al^{VI} = 0.15-0.32$  a.p.f.u.,  $X_{Mg} = 0.65-0.72$ ) is locally observed. Amphibole

404 porphyroclasts are wrapped around by the pervasive mylonitic foliation (Fig. 2b,c), mainly defined by  
405 phengitic white mica (Si = 3.30-3.37 a.p.f.u.) locally rimmed by muscovite (Si = 3.20-3.30 a.p.f.u.),  
406 associated with chlorite ( $X_{Mg} = 0.65-0.67$ ), quartz, albite ( $Ab_{96-100}$ ) and epidote ( $Ps_{20-26}$ ). Titanite, allanite and  
407 minor rutile occur as accessory minerals. Titanite is present both as large (100-200  $\mu m$  in size) subhedral  
408 grains with common opaque intergrowths, interpreted as relics of the magmatic protolith, and as small (<  
409 10  $\mu m$ ) euhedral grains aligned in the main foliation, interpreted as metamorphic (Fig. 4a). Due to the very  
410 small size of the metamorphic titanite, only the larger magmatic grains have been dated.

411 Sample 06-17 (N36°17'33.0" E76°34'58.2" - 3870m) is a poorly deformed, hornblende + biotite-  
412 bearing meta-diorite, still preserving the porphyric structure of the protolith (Fig. 2d). The deep-green, mm-  
413 sized, magmatic hornblende ( $Amp_1$ : magmatic Fe-hornblende; Si = 6.5-7.3 a.p.f.u.,  $Al^{IV} = 0.7-1.5$  a.p.f.u.,  $Al^{VI}$   
414 = 0.3-0.6 a.p.f.u.,  $X_{Mg} = 0.34-0.50$ ) is surrounded by a light green actinolitic rim ( $Amp_2$ : metamorphic  
415 actinolite; Si = 7.5-7.8,  $Al^{IV} = 0.2-0.4$  a.p.f.u.,  $Al^{VI} = 0.1-0.3$  a.p.f.u.,  $X_{Mg} = 0.54-0.56$ ), in turn overgrown by a  
416 thin outermost rim of deep-green Fe-hornblende ( $Amp_3$ : metamorphic Fe-hornblende; Si = 6.8-7.1 a.p.f.u.,  
417  $Al^{IV} = 0.9-1.2$  a.p.f.u.,  $Al^{VI} = 0.4-0.6$  a.p.f.u.,  $X_{Mg} = 0.38-0.45$ ) (Fig. 2e,f). The former plagioclase phenocrysts of  
418 the protolith are replaced by mm-sized, slightly zoned albite porphyroblasts (core:  $Ab_{97-100}$ ; rim:  $Ab_{95-97}$ )  $\pm$   
419 epidote ( $Ps_{25-30}$ ), whereas magmatic biotite is replaced by fine-grained aggregates of greenish-brown biotite  
420 ( $X_{Mg} = 0.40-0.42$ ; Ti = 0.12-0.13 a.p.f.u.) + epidote ( $Ps_{15-20}$ ) + minor chlorite ( $X_{Mg} = 0.45-0.47$ ) (Fig. 2f).  
421 Amphibole and albite porphyroblasts are set in a fine-grained matrix consisting of epidote + albite + minor  
422 quartz. Titanite occurs as large (up to 1 mm) aggregates of fine-grained euhedral crystals (Fig. 4b-d),  
423 interpreted as relics of the magmatic protolith.

424 The observed mineral assemblages and compositions suggest that both samples 06-10 and 06-17  
425 preserve the evidence of two distinct metamorphic events. The first event was more pervasive and was  
426 responsible for the growth of the actinolitic rim at the expenses of the magmatic hornblende in both  
427 samples, in equilibrium with albite + epidote + chlorite,  $\pm$  phengite (sample 06-10),  $\pm$  biotite (sample 06-17);  
428 the second event is marked by the appearance of the hornblende outermost rim in equilibrium with albite  
429 + epidote,  $\pm$  muscovite and chlorite in sample 06-10, and  $\pm$  biotite in sample 06-17. Equilibrium P-T  
430 conditions of the first metamorphic event were calculated for the assemblages Act + Ab + Phe + Chl + Ep +  
431 Qtz + Rt + Ttn of sample 06-10, which resulted to be  $482 \pm 20$  °C,  $11.5 \pm 1.2$  kbar (i.e. transition between  
432 greenschist- and blueschist-facies conditions; Fig. SM1); the second metamorphic event was constrained at  
433  $512 \pm 30$  °C,  $4.5 \pm 1.7$  kbar (i.e. transition between greenschist- and amphibolite-facies conditions; Fig. SM1)  
434 using the Hbl + Ab + Mu + Chl + Ep + Qtz + Rt + Ttn assemblage of sample 06-10 (Table 2). Mineral  
435 assemblages of sample 06-17 do not define enough reactions for Average PT to work, but are nevertheless  
436 consistent with the results obtained from sample 06-10. Overall, these data suggest a clockwise P-T  
437 evolution characterized by relatively high-P peak conditions of  $\sim 480$ °C, 11 kbar followed by decompression  
438 coupled with moderate heating at  $\sim 510$ °C, 4.5 kbar.



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### 5.1.2.2 Two-micas, garnet-bearing, metapelite (sample 06-115)

At its uppermost structural level, the Surukwat Complex mostly consists of metapelitic lithologies with minor intercalations of biotite-rich amphibolites and impure marbles likely derived from marl and limestone protoliths, respectively (Fig. 1d). Two of these metapelites, among which sample 06-115 studied in this work, have been petrologically investigated by Groppo and Rolfo (2008). Sample 06-115 is a two-micas, garnet-bearing micaschist consisting of quartz, muscovite, biotite, garnet, plagioclase ( $An_{5-18}$ ), chlorite and accessory ilmenite (Fig. 2g). The main schistosity ( $S_m$ ) is defined by the alignment of muscovite ( $Si = 3.00 - 3.10$  a.p.f.u.), biotite ( $X_{Mg} = 0.42-0.50$ ;  $Ti = 0.09-0.11$  a.p.f.u.) and ilmenite, and derives from the transposition of an earlier foliation ( $S_{m-1}$ ) defined by the same phases and still preserved in the microlithons (Fig. 2g). Garnet porphyroblasts ( $Alm_{71-75}Sp_{12-15}Prp_{7-9}Grs_{2-3}$ ) are enveloped by the  $S_m$  and overgrow the  $S_{m-1}$ , still preserved as an internal foliation (Fig. 2g). The outermost garnet rim is characterized by a sharp increase in  $X_{Mn}$  ( $Alm_{69-71}Sp_{20-21}Prp_{6-7}Grs_{2-3}$ ), likely reflecting diffusional re-equilibration at the onset of the  $S_m$  development. Late chlorite flakes ( $X_{Mg} = 0.45-0.47$ ) statically overgrow the  $S_m$ .

Monazite and rare xenotime are present as accessory minerals. Monazite occurs as subhedral grains within garnet and as elongate grains and clusters of grains aligned parallel to the dominant foliation in the matrix (Fig. 5).

The results of thermodynamic modeling (pseudosection approach: Groppo and Rolfo, 2008) suggest that the peak assemblage (Grt + Wm + Bt + Pl + Qz + Ilm) grew at 580-600 °C, 8-9 kbar; consistent results are given by the Average PT method applied on the same assemblage ( $645 \pm 26$  °C,  $8.2 \pm 1.2$  kbar) (Table 2; Fig. SM1). Basing on the pseudosection results, Groppo and Rolfo, 2008 further constrained the main foliation development at ca. 500 °C, 5 kbar. Furthermore, combining the results obtained from sample 06-115 with those obtained from a second metapelite sample (06-117), the same authors inferred a steep and narrow anticlockwise P-T evolution for this portion of the Surukwat Complex (Fig. SM1).

### 5.1.3 Aghil and Sughet magmatic bodies (samples 06-26, 06-108)

Sample 06-26 (N36°11'02.4" E76°37'32.7" - 4905 m) was collected from the Aghil Granodiorite body, in the proximity of the Aghil Pass. It is a biotite-bearing porphyritic granite crosscutting the main granodiorite, with perthitic K-feldspar crystals up to several centimeters in length (Fig. 2h). Plagioclase is zoned, with the Ca-richer core locally altered in sericite  $\pm$  saussurite, and a thin albite rim. Biotite is partially replaced by chlorite. Zircon and apatite are abundant among the accessory minerals and are often included in biotite.

Sample 06-108 (N36°04'08.3" E76°24'52.4" - 3887 m) was collected from the Sughet Granodiorite body, near Sughet Jagal. It is a biotite-bearing granodiorite with poikilitic K-feldspar, zoned plagioclase partially altered in sericite, and brown biotite pervasively replaced by chlorite (Fig. 2i); apatite, zircon, and allanite occur as accessory minerals.

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## 5.2 Geochronology results

### 5.2.1 Zircon

Zircon from the Aghil (06-26) and Sughet (06-108) granodiorite bodies is euhedral with well-developed oscillatory zoning (Fig. 6). Sample 06-26 yielded reproducible ages and consistent REE patterns (Fig. 7a). Eight analyses define a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $83 \pm 1$  Ma (MSWD=1.6) (Fig. 7a and Table 3). Results from sample 06-108 exhibit more scatter with one Proterozoic and 11 Cretaceous apparent ages. The Proterozoic age is interpreted as a xenocrystic core. Seven of the younger ages define a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $102 \pm 1$  Ma (MSWD=0.9; Fig. 7b, Table 3). Two slightly older analyses are interpreted to reflect mixture of xenocrystic cores and younger magmatic zircon. The two younger analyses are inferred to record younger disturbance possibly associated with emplacement of the latest magmatic products of the Sughet granodiorite suite.

### 5.2.2 Titanite

Titanite  $^{206}\text{Pb}/^{238}\text{U}$  analyses from samples 06-10 and 06-17 (Surukwat Complex) generally contain low U (average 2 ppm U) and high common Pb resulting in mixing array on Tera-Wasserburg plots (Fig. 8). Regression of analyses from sample 06-17 yields a well-defined 3-D isochron intercept age of  $796 \pm 29$  Ma (MSWD=1.7), that is consistent with a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $772 \pm 31$  Ma (MSWD=2.4) for the same data. All analyses from sample 06-10 (large titanite grains) are dominated by common Pb, plotting near the upper intercept, with a regression that defines a lower intercept of  $635 \pm 410$  Ma, consistent with results of sample 06-17. Titanite ages from both samples are interpreted to record the protolith age of the orthogneiss bodies.

### 5.2.3 Xenotime and monazite

Xenotime analyses from sample 06-1 (Bazar Dara Slates Unit) define a weighted mean age of  $174 \pm 11$  Ma (MSWD = 2.8) (Fig. 9a, Table 6) and show a middle REE (MREE) enriched patterns (Fig. 9a) with strong depletion in light REE (LREE) characteristic of metamorphic xenotime (e.g. Aleinikoff et al., 2015). The xenotime grains occur in the matrix and have an ambiguous relationship with the dominant foliation. The xenotime ages are interpreted to record metamorphism at ca. 170 Ma. Two matrix monazite grains in sample 06-1 yield  $^{206}\text{Pb}/^{238}\text{U}$  ages of  $153 \pm 3$  Ma and  $161 \pm 4$  Ma, which we interpret to be consistent with the ca. 170 Ma metamorphic age derived from xenotime, whereas a much younger metamorphic age of  $35 \pm 17$  Ma is recorded by a single matrix monazite.

In sample 06-115 (Surukwat Complex), xenotime is very rare. One grain of xenotime gave a  $^{204}\text{Pb}$  corrected  $^{206}\text{Pb}/^{238}\text{U}$  age of  $165 \pm 6$  Ma, which is consistent with the ca. 170 Ma metamorphic age derived

509 from xenotime in sample 06-1. Monazite in sample 06-115 dominantly occurs in the matrix but is locally  
510 preserved within garnet (Fig. 5). The monazite shows negative Eu anomalies and heavy REE depletion (Fig.  
511 9b) consistent with growth in the presence of feldspar and garnet, respectively (e.g. Rubatto et al., 2013).  
512 Two monazite  $^{206}\text{Pb}/^{238}\text{U}$  age populations occur in this sample: (i) two monazite grains included in garnet  
513 gave  $^{206}\text{Pb}/^{238}\text{U}$  ages of  $144 \pm 4$  Ma and  $157 \pm 6$  Ma (Fig. 9b), thus defining a ca. 150 Ma monazite  
514 generation; (ii) nine analyses from matrix monazite in sample 06-115 (Fig. 9b, Table 5) define a weighted  
515 mean age of  $66 \pm 2$  Ma (MSWD = 1.1) (second monazite generation). The remaining grains record ages  
516 between 108 and 70 Ma that are interpreted to reflect partial resetting of the older monazite formed  
517 originally at ca. 150 Ma. This interpretation is supported by core to rim Th zoning observed in grains that  
518 give U/Pb ages  $>70$  Ma (Fig. 5).

## 6. Discussion

### 6.1 Pre-Cenozoic geodynamic evolution of the Aghil Range

#### 6.1.1 The Surukwat Complex: a relic of Neoproterozoic basement with a possible late Palaeozoic cover

525 The U-Pb titanite ages obtained for the meta-granodiorite and meta-diorite (samples 06-10 and 06-17)  
526 from the Surukwat Complex tightly constrain the formation of the igneous protoliths at  $796 \pm 29$  Ma, i.e.  
527 during the Neoproterozoic.

528 Although not common, fragments of Neoproterozoic crystalline basement have been reported from  
529 different terranes in the central Tibetan Plateau (see Zhu et al., 2013 for a review). In the central Lhasa and  
530 Amdo terranes, U-Pb dating of zircons in gneissic rocks yielded 787-748 Ma (Nam Tso area: Hu et al., 2005)  
531 and 920-820 Ma (Amdo: Guynn et al., 2006, 2012) ages, respectively. Pre-Cambrian basement exposures  
532 are not reported so far from the Qiangtang terrane, the oldest basement rocks having been dated as  
533 Ordovician (Pullen et al., 2011; Zhao et al., 2014) or Cambro-Ordovician (Kapp et al., 2000). However, the  
534 existence of a  $> 740$  Ma old basement in the Qiangtang terrane is suggested by Neoproterozoic inherited  
535 zircon ages obtained from an Early-Cretaceous granite from the Longmu Co area, western Tibet (Leloup et  
536 al., 2012). Neoproterozoic ages (ca. 865-825 Ma) have been also obtained for fragments of crystalline  
537 basement exposed in the southern part of the Songpan-Ganze terrane, in eastern Tibet (Roger and  
538 Calassou, 1997; Zhou et al., 2002, 2006a,b). Relicts of a possibly pre-Cambrian basement are also reported  
539 from the Karakorum (e.g. Le Fort et al., 1994; Zanchi and Gaetani, 2011), South Pamir (East Hindu Kush)  
540 (e.g. Zanchi and Gaetani, 2011) and Central Pamir (e.g. Schwab et al., 2004) terranes of the Pamir-  
541 Karakorum Range. All these data suggest that the different Gondwana-derived terranes which constitute  
542 the Tibetan Plateau and the Pamir-Karakorum Range are characterized by similar pre-Cambrian igneous  
543 basements, likely formed during the initial stages of Rodinia break-up (e.g. Guynn et al., 2012 and

544 references therein). Our data allow the recognition of a new exposure of such Pre-Cambrian basement in  
545 the Surukwat Complex: this finding represents the westernmost occurrence, known so far, of Pre-Cambrian  
546 basement rocks in the Tibetan Plateau.

547 The Neoproterozoic basement of the Surukwat Complex is overlaid by a meta-sedimentary cover  
548 whose protoliths (i.e. conglomerate immediately overlying the igneous basement followed upward by  
549 sandstone and then mudstone with marl and limestone intercalations) are compatible with deposition in a  
550 passive continental margin setting (Fig. 1c). The age of these sedimentary protoliths is unknown, but it  
551 should be not younger than Triassic, because our data demonstrate that during Middle Jurassic these  
552 sediments were already metamorphosed (see below, Section 6.1.2). Basing on the strong similarities with  
553 late Palaeozoic meta-sedimentary sequences overlying the crystalline basement in the central Lhasa, Amdo  
554 and Qiangtang terranes (Kapp et al., 2000, 2003b, 2007; Leier et al., 2007; Pullen et al., 2008; Guynn et al.,  
555 2012), we suggest that the protoliths of the meta-sedimentary sequence exposed in the upper portion of  
556 the Surukwat Complex might be late Palaeozoic in age. In the absence of geochronological data  
557 constraining the age of the sedimentary protoliths, however, this hypothesis remains speculative, but is  
558 nevertheless consistent with the observed lithological associations. Our results thus suggest that the  
559 Surukwat Complex is a coherent slice of Neoproterozoic crystalline basement of dioritic to granodioritic and  
560 granitic composition, with a possibly late Palaeozoic sedimentary cover (Fig. 1c) deposited on a passive  
561 continental margin during the Gondwana break-up.

### 33 6.1.2 The Middle Jurassic collision between the Bazar Dara Slates and the Surukwat Complex and the 34 Cimmerian Orogeny

35 The first tectono-metamorphic event experienced by the studied units is Middle Jurassic (ca. 170 Ma). This  
36 event is registered by xenotime and monazite in the two micas phyllite of the Bazar Dara Slates Unit  
37 (sample 06-1), and by rare xenotime grains in the garnet-bearing micaschist from the Surukwat Complex  
38 Unit (sample 06-115).

39 Evidence of a Late Triassic - Middle Jurassic orogenic event are widespread in both the Tibetan  
40 Plateau and the Pamir-Karakorum Range; this event, known as Cimmerian Orogeny (Sengör, 1984), led to  
41 the final amalgamation of the Cimmerian terranes (i.e. Qiangtang, Amdo, Central Pamir, South Pamir and  
42 Karakorum terranes) to the southern margin of Asia (i.e. Songpan-Ganze/Karakul Mazar and Kunlun  
43 terranes) through the complete closure of the Paleo-Tethys, and of the smaller Longmu Tso-Shuanghu,  
44 Rushan-Pshart and Wakan oceanic basins. The timing of each collision has been mostly determined basing  
45 on either stratigraphic or magmatic constraints, whereas metamorphic constraints have been less  
46 frequently used. Previous studies have shown that the final collision between each terrane occurred at  
47 slightly different ages from west to east. In the central Tibetan Plateau, the final collision of the Qiangtang  
48 terrane with the Songpan-Ganze and Kunlun terranes occurred during the Late Triassic - Early Jurassic (e.g.  
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579 Dewey et al., 1988; Matte et al., 1996; Roger et al., 2003, 2010; Dai et al., 2013; Cao et al., 2015; Liu et al.,  
580 2015), and the amalgamation of South Qiangtang and North Qiangtang was broadly coeval, independently  
581 from the model chosen to explain the nature of the Central Qiangtang Metamorphic Belt (i.e. underthrust  
582 model vs intra-Qiangtang suture model, see section 2). In the Pamir-Karakorum Range the accretions of  
583 Central Pamir to Karakul Mazar (North Pamir) and of South Pamir to Central Pamir were broadly coeval at  
584 ca. 200 Ma, i.e. during the latest Triassic - earliest Jurassic (e.g. Gaetani et al., 1993; Zanchi et al., 2000,  
585 2012; Zanchi and Gaetani, 2011; Angiolini et al., 2013; Robinson et al., 2015; Zanchetta et al., 2018),  
586 whereas the collision of Karakoram with South Pamir is supposed to be Late Early Jurassic (e.g. Searle and  
587 Tirrull, 1991; Gaetani et al., 1993; Angiolini et al., 2013).

588 In this framework, we therefore interpret the Middle Jurassic xenotime and monazite ages  
589 recorded by the two micas phyllite of the Bazar Dara Slates Unit (sample 06-1) to be related to the  
590 Cimmerian orogenic event, i.e. to the collision between the Bazar Dara Slates and the Surukwat Complex  
591 (Fig. 10); according to the petrologic results, this collisional event was responsible for a greenschist-facies  
592 (i.e. ~320°C, 5 kbar) metamorphic overprint in the Bazar Dara Slates Unit (Fig. 11a,b). This is consistent with  
593 the stability field of both xenotime and monazite, that can grow in low-grade pelitic schists (Pan, 1997;  
594 Wing et al., 2003; Bollinger and Janots, 2006; Janots et al., 2006; Rasmussen et al., 2007; Krenn and Finger,  
595 2007).

596 Due to the scarcity of xenotime in the metapelite of the Surukwat Complex (sample 06-115), the  
597 interpretation of its age and of the P-T conditions of its growth remain uncertain. It has been demonstrated  
598 that xenotime remains stable with increasing metamorphic grade in garnet-absent rocks, whereas it reacts  
599 out continuously as garnet grows, and is typically gone from the matrix assemblage in samples from the  
600 middle garnet zone (Spear and Pyle, 2002; Fitsimons et al., 2005). This constrains the growth of xenotime at  
601 temperatures below the first appearance of garnet (i.e.  $T < 500^{\circ}\text{C}$  according to the pseudosection modelling  
602 in Groppo and Rolfo, 2008) and explains why xenotime is so scarce in the garnet-bearing sample 06-115.  
603 The ca. 170 Ma age obtained from the rare xenotime grains in this sample might be thus related to the  
604 same Cimmerian orogenic event registered by the Bazar Dara Slates Unit. An alternative hypothesis could  
605 be to correlate the ca. 170 Ma age to the earliest stage of development of an accretionary complex on the  
606 southern margin of the Surukwat Complex (see below, section 5.1.3; Fig. 11a).

### 6.1.3 *The Late Jurassic formation of an accretionary prism on the southern margin of the Surukwat Complex*

609 The Surukwat Complex experienced another significant tectono-metamorphic event during the Late  
610 Jurassic at ca. 150 Ma, responsible for the growth of a first generation of monazite in the metapelite  
611 sample 06-115 (two grains:  $157 \pm 6$  Ma and  $144 \pm 4$  Ma). Microstructural observations show that this  
612 monazite generation is preserved in garnet, whereas in the matrix it is partially reset by the growth of a  
613 younger monazite generation (Fig. 5). It has been demonstrated that the first appearance of monazite in

614 metapelitic rocks is strongly influenced by the bulk rock composition, particularly by the CaO content (e.g.  
615 Foster and Parrish, 2003; Wing et al., 2003) and the CaO/Na<sub>2</sub>O ratio (e.g. Janots et al., 2008), and to a lesser  
616 extent by the Al<sub>2</sub>O<sub>3</sub> content and FeO/MgO ratio (e.g. Fitsimons et al., 2005). Janots et al. (2008)  
617 demonstrated that in metapelites with low Ca/Na ratios (i.e. CaO/Na<sub>2</sub>O<0.54), monazite is formed through  
618 the complete breakdown of allanite at T > 580°C, whereas higher Ca/Na ratios enhance the allanite stability  
619 towards higher temperatures. The CaO/Na<sub>2</sub>O ratio of sample 06-115 is 0.26 (Groppo and Rolfo, 2008),  
620 therefore it is likely that in this sample monazite grew at T > 580°C, through a reaction such as that  
621 proposed by Janots et al. (2008) (i.e. allanite + apatite + Al-Fe-Mg phases<sub>1</sub> → monazite + anorthite + Al-Fe-  
622 Mg phases<sub>2</sub>, with chlorite and garnet involved as Al-Fe-Mg phase<sub>1</sub> and phase<sub>2</sub>, respectively). Garnet  
623 growth was modelled at ~600 °C, 9 kbar (i.e. at peak conditions; Groppo and Rolfo, 2008), therefore we  
624 suggest that the ca. 150 Ma registered by monazite grains included in garnet might be associated to the  
625 peak metamorphic event, and that monazite formation was nearly coeval with garnet growth (Fig. 11d), as  
626 also evidenced by its trace element pattern (Fig. 9b).

627 To interpret the geodynamic significance of this Late Jurassic metamorphic event, the apparently  
628 contrasting anticlockwise and clockwise P-T evolutions inferred for samples 06-115 and 06-10/17 should be  
629 considered (Groppo and Rolfo, 2008). Anticlockwise P-T paths at relatively high P-T conditions are  
630 characteristic of accretionary systems during the early stages of underflow, as suggested for the Franciscan  
631 melange (e.g. Platt, 1975; Cloos, 1985, 1986; Ernst, 1988; Krogh et al., 1994), the Coastal Cordillera of  
632 south-central Chile (e.g. Willner et al., 2004; Willner, 2005; Hyppolito et al., 2014) and, possibly, the North  
633 America Cordillera (Perchuck et al., 1999). In such geodynamic setting, rocks that were piled up earlier to  
634 the hot hanging wall of the subduction channel in the deepest part of the accretionary complex, were  
635 metamorphosed at higher temperatures for a given pressure (typically at albite-epidote amphibolite-facies  
636 conditions, transitional between high-pressure greenschist facies and eclogite facies; ~600-650 °C, 8-12  
637 kbar), compared to material subducted later (Perchuck et al., 1999). In the following accretionary stages,  
638 the hanging wall became progressively cooler and isotherms were continuously displaced to greater depth  
639 due to the continuous de-hydration of the later subducted material, thus implying that the early subducted  
640 rocks were exhumed at temperatures lower than those experienced during subduction, whereas the rocks  
641 involved later in the accretionary process recorded a normal clockwise P-T path (e.g. Willner et al., 2004;  
642 Willner, 2005; Hyppolito et al., 2014). Anticlockwise P-T metamorphic evolutions related to the inception of  
643 subduction are not commonly observed because of the extremely small areas affected by this type of  
644 metamorphism (≤ tens of km<sup>2</sup>, with thickness of few hundred meters: Wakabayashi, 2004). P-T conditions  
645 and evolutions recorded by the studied samples fit well with this geodynamic scenario (Fig. 10e-f, 11c-f).  
646 Moreover, it has been recently demonstrated that in similar tectonic settings, the accretionary process did  
647 not result in a chaotic mixing, but rather in the formation of a “pseudo”-coherent unit (Hyppolito et al.,

648 2014). This is also consistent with our observation that the Surukwat Complex consists of a tectonic  
649 intermingling of lithological units which nevertheless maintain their internal coherence.

650 To summarize, we suggest that during the Late Jurassic – Early/Middle Cretaceous, an accretionary  
651 prism developed on the southern margin of the Surukwat Complex (Fig. 10e-f). The first material involved in  
652 the subduction was a portion of the upper Palaeozoic (?) sedimentary cover (i.e. sample 06-115), which was  
653 subducted in a relatively hot environment at ca. 150 Ma (Fig. 11c,d). The successive subduction of the  
654 thinned southern portion of the pre-Cambrian basement (i.e. samples 06-10 and 06-17) occurred in a  
655 cooler environment, during the early exhumation of the firstly subducted metasedimentary rocks (Fig.  
656 11e,f).

#### 658 *6.1.4 The Late Cretaceous collision between the Surukwat Complex and the Lhasa terrane*

659 The Surukwat Complex registered the last tectono-metamorphic event during the Late Cretaceous: this  
660 event is recorded in the metapelite sample 06-115 by the growth of monazite at  $66 \pm 2$  Ma. This second  
661 generation of monazite is ubiquitous in the matrix and appears in equilibrium with the main foliation,  
662 whose P-T conditions of formation were constrained by Groppo and Rolfo (2008) at  $\sim 500^\circ\text{C}$ , 5 kbar.

663 Evidence of Late Cretaceous metamorphism has been reported from western Tibet and Pamir-  
664 Karakorum Range and are interpreted as the result of collision between the Lhasa and South Qiangtang  
665 terranes (western Tibet) and between the Kohistan-Ladakh and Karakorum terranes (Pamir-Karakorum),  
666 respectively. More specifically, in western Tibet, tight constraints on the time of this collision have been  
667 recently presented by Liu L. et al. (2017) and Liu D. et al. (2017) basing on the Late Cretaceous age of syn-  
668 orogenic magmatism. In the Pamir-Karakorum Range, the timing of collision between Kohistan-Ladakh and  
669 Karakorum has been constrained as Middle to Late Cretaceous basing on stratigraphic data (e.g. Gaetani et  
670 al., 1990b, 1993; Robinson et al., 2004; Zanchi and Gaetani, 2011; Gaetani, 2016), age of syn-orogenic  
671 magmatism (e.g. Fraser et al., 2001; Heuberger et al., 2007; Searle and Philipps, 2007; Searle, 2011) and  
672 metamorphic studies (e.g. Fraser et al., 2001; Streule et al., 2009; Searle, 2011). A complex and long-lasting  
673 history of tectonic and magmatic activity associated with the Kohistan-Karakorum suture zone, possibly  
674 extending up to Eocene time, is also proposed by Heuberger et al. (2007).

675 The  $66 \pm 2$  Ma monazite age recorded by sample 06-115 can be therefore interpreted as the  
676 evidence of this collisional event (Fig. 10g, 11g-h). In this framework, the  $83 \pm 1$  Ma zircon age obtained  
677 from the porphyric granite of the Aghil Granodiorite body (sample 06-26) is perfectly compatible with a  
678 subduction-related syn-orogenic magmatism preceding the final collision between the Surukwat Complex  
679 and the Lhasa terrane (Fig. 11g). It is worth noting that Liu L. et al. (2017) obtained very similar results (78-  
680 80 Ma) from granitic rocks collected from the same area. A similar interpretation can be extended also to  
681 the  $102 \pm 1$  Ma zircon age obtained from the Sughet Granodiorite (sample 06-108): however, the Sughet

682 Granodiorite intrusive body is located south of the Karakorum fault and it is therefore not directly  
683 correlated to the geodynamic history of the Surukwat Complex (see below, section 5.2).

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## 685 **6.2 Solving the Tibet-Pamir-Karakorum geopuzzle**

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687 The new field, petrological and geochronological data presented in this paper provide new insights toward  
688 a better understanding of the Tibet-Pamir-Karakorum geopuzzle, allowing to correlate the different  
689 terranes of Central Tibet with those of the Pamir-Karakorum Range on both sides of the Karakorum fault.

690 We propose that (Fig. 12):

691 (1) the Bazar Dara Slates Unit might be the equivalent of the Sonpan-Ganze terrane of central Tibet, as  
692 already suggested by other authors (e.g. Gaetani et al., 1990a, 1991; Matte et al., 1996; Gaetani, 1997;  
693 Schwab et al., 2004; Rolfo et al., 2014) and can be correlated to the Karakul-Mazar terrane of the  
694 Pamir-Karakorum Range. Lithological and geochronological data from the Karakul-Mazar terrane of the  
695 eastern portion of Northern Pamir are especially similar to those described here for the Bazar Dara  
696 Slates Unit. These similarities include: (i) the occurrence of Triassic granitoids intruded in  
697 metasedimentary lithologies in both terranes (Robinson et al. 2007; Liu et al., 2015); (ii) the age of  
698 metamorphism (ranging from greenschist- to amphibolite-facies conditions) was constrained to be  
699 Early to Middle Jurassic (160-200 Ma) in the schists exposed along the Muztaghata massif of eastern  
700 Northern Pamir (Robinson et al., 2012) and is remarkably similar to the Middle Jurassic age of 170 Ma  
701 registered by the studied sample 06-1, which also experienced greenschist-facies metamorphism.  
702 Robinson et al. (2012) interpreted this age as the timing of the collision between the Karakul-Mazar  
703 terrane and the Central Pamir during final closure of the Paleo-Tethys Ocean, as well as we interpret  
704 the Middle Jurassic age as dating the collision between the Bazar Dara Slates and the Surukwat  
705 Complex.

706 (2) the Surukwat Complex might be correlated to the southern margin of the South Qiangtang terrane of  
707 central Tibet (see also Gaetani et al., 1990a, 1991; Rolfo et al., 2014), and shows significant similarities  
708 with the tectono-stratigraphy and metamorphic evolution of the Amdo terrane. These similarities  
709 include: (i) the occurrence of a Neoproterozoic basement in both the Amdo and Surukwat Complex. In  
710 both cases this basement consists of felsic to intermediate gneisses derived from igneous protoliths  
711 and of a metasedimentary cover derived from late Palaeozoic sediments deposited in a passive margin  
712 setting (Guynn et al., 2006, 2012); (ii) widespread evidence of Jurassic metamorphism preceding the  
713 Lhasa-Qiangtang collision. Peak metamorphic conditions in the Amdo terrane vary from ~600°C, 8 kbar  
714 (remarkably similar to peak conditions estimated for sample 06-115) to ~700°C, 10 kbar, and  
715 metamorphic peak has been dated as Early Jurassic (ca. 178 Ma: Guynn et al., 2006, 2013). The Amdo  
716 terrane thus experienced peak metamorphism earlier than the Surukwat Complex (ca. 150 Ma),

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717 consistently with a diachronous activity of the Bangong-Nunjiang suture zone from east to west (Fig.  
718 12d-f).

719 Opposite to what preliminary suggested in Groppo and Rolfo (2008), the Surukwat Complex cannot be  
720 correlated to the Central Qiangtang Metamorphic Belt (CQMB), for three main reasons: (i) the CQMB  
721 mostly consists of a tectonic melange (e.g. Kapp et al., 2000, 2003a), whereas the Surukwat Complex is  
722 a coherent portion of an old basement with its original sedimentary cover still preserved; (ii) peak  
723 metamorphic conditions in some blocks of the CQMB reached blueschist and eclogite facies conditions  
724 (e.g. Kapp et al., 2000, 2003a; Zhang et al., 2006a,b, 2011; Pullen et al., 2008, 2011; Zhang and Tang,  
725 2009; Zhai et al., 2011b; Zhao et al., 2014), thus reflecting a significantly different geothermal gradient  
726 with respect to the Surukwat Complex; (iii) the timing of metamorphism in the CQMB is Middle to Late  
727 Triassic (244-223 Ma: Pullen et al., 2008, 2011) whereas in the Surukwat Complex it is Late Jurassic (ca.  
728 150 Ma).

729 (3) From a paleogeographic point of view, the Surukwat Complex occupies an intermediate position  
730 between the South Qiangtang terrane and the South Pamir terrane (e.g. Robinson, 2015, Chapman et  
731 al., 2018 and references therein), preserving different margins of the same micro-plate. The  
732 fundamental difference between the Surukwat Complex and South Pamir is that the latter does not  
733 show the evidence of Cretaceous metamorphism, having been “protected” from the collision with the  
734 Kohistan-Ladakh terrane by the interposed Karakorum terrane (Fig. 12g). Instead, the Karakorum  
735 terrane registered an important metamorphic and deformational event during the Late Cretaceous  
736 (between 83 and 62 Ma), that was interpreted as due to the collision with the Kohistan-Ladakh Arc  
737 (Fraser et al., 2001).

738 (4) In this framework, we propose that the non-metamorphic Shaksgam Sedimentary Belt and the Sughet  
739 Granodiorite body, now tectonically interposed between the Surukwat Complex and the Karakorum  
740 terrane, do not belong neither to the Surukwat Complex nor to the Karakorum terrane, but instead  
741 they are part of the South Pamir terrane (see also Robinson, 2009 and Gaetani and Leven, 2014). In  
742 other words, their actual location would not reflect their original position in pre-Cenozoic times, but it  
743 would be related to the offset effects of the Karakorum fault. This hypothesis is supported by both  
744 petrologic and geochronologic data showing that the Shaksgam Sedimentary Belt escaped the Late  
745 Cretaceous metamorphic and deformational event experienced by the Surukwat Complex and likely  
746 related to the collision with the Lhasa Terrane (Fig. 10g, 12g). Moreover, the  $102 \pm 1$  Ma zircon age  
747 obtained from the Sughet Granodiorite body is perfectly compatible with the Late Cretaceous  
748 calcalkaline subduction-related magmatism responsible for the emplacement of large intrusive bodies  
749 in South Pamir as well as in the North Karakorum terrane (e.g. Karakorum Batholith, 106-95 Ma)  
750 (Debon et al., 1987; Fraser et al., 2001; Schwab et al., 2004; Zanchi and Gaetani, 2011), which has been  
751 related to a north-directed low-dipping subduction below Karakorum.

752 **Supplementary material**

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753 Supplementary data to this article including (i) phase diagrams supporting the P-T evolution constrained for  
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754 the studied samples (Fig. SM1) and (ii) the complete set of U-Pb geochronologic data for zircon (Table  
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755 SM1a), titanite (Table SM1b), monazite (Table SM1c) and xenotime (Table SM1d) can be found online at:  
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## FIGURE CAPTIONS

**Fig. 1 - (a)** Simplified tectonic map of the central-western Tibetan Plateau and Pamir-Karakorum Range (modified after Gaetani et al., 1991 and Robinson, 2009), with location of the Aghil Range (black rectangle). From north to south, the main terranes and sutures of the Tibetan Plateau are TB: Tarim Basin; KL: Kunlun; SG: Songpan-Ganze; NQ: North Qiangtang; SQ: South Qiangtang; LH: Lhasa; I: India; KS: Kunlun Suture; JS: Jinsha Suture; LSSZ: Longmu Tso-Shuanghu Suture Zone; BNSZ: Bangong-Nunjiang Suture Zone; ITSZ: Indus-Tsangpo Suture Zone, whereas those of the Pamir- Karakorum Range are KM: Karakul-Mazar (North Pamir); CP: Central Pamir; SP: South Pamir; KK: Karakorum; KH/LK: Kohistan-Ladakh; TS: Tanyamas Suture; RPSZ: Rushan-Pshart Suture Zone; TBZ: Tirich Boundary Zone; SSZ: Shyok Suture Zone. KKF: Karakorum Fault; KF: Karakax Fault; LCF/ATF: Longmu Co Fault / Althyn Tagh Fault. **(b)** Geologic map of the Aghil Range, between Kunlun (Yarkand River) and Karakorum (Shaksgam River), modified after Gaetani et al. (1991) and Rolfo et al. (2014), with location of the studied samples (black stars). **(c)** Sketch of the pre-Mesozoic original relationships between the crystalline basement and the sedimentary cover of the Surukwat Complex. **(d)** Geologic cross-section from Ilik to Sughet Jagal (A-B in (b)), with location of the studied samples (white stars).

**Fig. 2 – Representative microstructures of the studied samples. Bazar Dara Slates Unit - Sample 06-1: (a)** Two-micas + chlorite phyllite: the main foliation  $S_m$ , defined by white mica + chlorite + biotite + ilmenite, is locally pervasively crenulated with the appearance of an  $S_{m+1}$  defined by white mica + ilmenite. Plane Polarized Light (PPL). *Surukwat Complex – Sample 06-10: (b)* The mylonitic foliation, defined by phengite, wraps around pluri-mm amphibole porphyroclasts (PPL). **(c)** Detail of an amphibole porphyroclast: the brownish-green amphibole core ( $Amp_1$ ) is a relic of the igneous protolith and it is rimmed by a green metamorphic actinolite ( $Amp_2$ ) (PPL). *Sample 07-17: (d)* The porphyric structure of the dioritic protolith is still preserved (note the mm-sized amphibole porphyroblasts set in a fine-grained matrix of albite + epidote + biotite ± chlorite) (PPL). **(e, f)** Details of strongly zoned amphibole crystals, with a dark green core ( $Amp_1$ , magmatic hornblende), a light green rim ( $Amp_2$ , metamorphic actinolite) and a discontinuous deep green outermost rim ( $Amp_3$ , metamorphic hornblende) (PPL). *Sample 06-115: (g)* The main foliation  $S_m$ , defined by white mica + biotite, wraps around mm-sized garnet porphyroblasts. An earlier  $S_{m-1}$  is preserved in the microlithons and as an internal foliation within garnet. Chlorite porphyroblasts overgrows the  $S_m$  (PPL). *Plutonic bodies – Sample 06-26: (h)* This porphyritic granite is characterized by cm-sized perthitic K-feldspar; biotite is fresh and plagioclase is slightly zoned (Crossed Polarized Light: XPL). *Sample 06-108: (i)* In this biotite-bearing granodiorite, K-feldspar is poikilitic and biotite is mostly replaced by chlorite (XPL).

1203 **Fig. 3** – BSE images of monazite (**a, b**) and xenotime (**c, d**) in sample 06-1 showing location of U/Pb-TE  
1204 analysis. Scale bar in all images is 50  $\mu\text{m}$ .

1205  
1206 **Fig. 4** – BSE images of titanite from samples 06-10 (**a**) and 06-17 (**b-d**) showing location of U/Pb analysis.

1207  
1208 **Fig. 5** – BSE images (**a-c**) and compositional maps (**d**) of monazite from sample 06-115 showing location of  
1209 U/Pb-TE analysis. Monazite grains in (a) are included within garnet and yield Jurassic U/Pb ages. Matrix  
1210 monazite grains (b-c) are Late Cretaceous in age. Matrix monazite that displays core to rim zoning in Th (see  
1211 4-2.2 and 6-1.1 in d) gives older U/Pb ages interpreted to reflect incomplete recrystallization of Jurassic  
1212 monazite.

1213  
1214 **Fig. 6** – Cathodoluminescence (CL) images of zircons from samples 06-108 (**a**) and 06-26 (**b**) showing  
1215 location of U/Pb-TE analysis (large circle) and additional TE analysis (small circle).

1216  
1217 **Fig. 7** – Tera-Wasserburg plots of U/Pb data and plots of chondrite normalized trace element data from  
1218 zircon in samples 06-26 (**a**) and 06-108 (**b**). Black ellipses and black symbols in chondrite normalized plots  
1219 are used in age interpretation; gray ellipses are not. Ellipses plotted at  $1\sigma$ . Age uncertainties reported at  
1220 95% confidence level (MSWD = mean square of weighted deviates). Tera-Wasserburg diagrams were made  
1221 using Isoplot (Ludwig, 2003). Chondrite normalized plots were calculated using values from McDonough  
1222 and Sun (1995).

1223  
1224 **Fig. 8** – Tera-Wasserburg plots of U/Pb data from titanite in samples 06-17 (**a**) and 06-10 (**b**). Ellipses are  
1225 plotted at  $1\sigma$ . Lower intercept age uncertainties reported at 95% confidence level (MSWD = mean square of  
1226 weighted deviates). Tera-Wasserburg diagrams were made using Isoplot (Ludwig, 2003).

1227  
1228 **Fig. 9** – Tera-Wasserburg plots of U/Pb data and plots of chondrite normalized trace element data from  
1229 xenotime, sample 06-1 (**a**) and monazite, sample 06-115 (**b**). Black ellipses and black symbols in chondrite  
1230 normalized plots are used in age interpretation; gray ellipses are not. Ellipses plotted at  $1\sigma$ . Analysis 5-2.1  
1231 is not plotted in (b) due large uncertainty in the  $^{207}\text{Pb}/^{206}\text{Pb}$  ratio. Age uncertainties reported at 95%  
1232 confidence level (MSWD = mean square of weighted deviates). Tera-Wasserburg diagrams were made  
1233 using Isoplot (Ludwig, 2003). Chondrite normalized plots were calculated using values from McDonough  
1234 and Sun (1995).

1235  
1236 **Fig. 10** – Pre-Cenozoic tectonic evolution of Western Tibetan Plateau as inferred from geochronological and  
1237 petrological data discussed in this paper. The sketch is especially focused on the evolution of the Surukwat



1238 Complex, whereas it is simplified for the other terranes. Abbreviations for the western Tibet terranes and  
1239 for the interposed oceans are: KL: Kunlun; BDs: Bazar Dara Slates Unit; SC: Surukwat Complex; LH: Lhasa; I:  
1240 India; *PT*: Paleo-Tethys; *BN*: Bangong-Nunjian Ocean; *NT*: Neo-Tethys. Abbreviations for suture zones are:  
1241 *KS*: Kunlun Suture; *JS*: Jinsha Suture; *BNSZ*: Bangong-Nunjian Suture Zone. The black rectangle refers to Fig.  
1242 11.

1243  
1244 **Fig. 11** – Enlargement of Fig. 10 focusing on the Mesozoic evolution of the Surukwat Complex and  
1245 explaining the development of the accretionary wedge on its southern margin (**a, c, e, g**). Colour codes are  
1246 the same as in Fig. 10. The P-T diagrams in (**b, d, f, h**) show the peak P-T conditions experienced by the  
1247 Bazar Dara Slates (b) and by the different thrust sheets of the Surukwat Complex (d, f, h) as inferred from  
1248 the Average PT results (with 1 $\sigma$  error ellipses). The P-T path of sample 06-115 is derived from Groppo and  
1249 Rolfo (2008), whereas that for sample 06-10 is tentatively inferred basing on the Average PT results.

1250  
1251 **Fig. 12** – (**a-h**) Simplified paleogeographic sketches showing the relative positions of the Cimmerian  
1252 Terranes in pre-Cenozoic times, as inferred from the data presented in this paper. The separation between  
1253 Central Pamir, South Pamir and Karakorum terranes in the Pamir-Karakorum Range is in agreement with  
1254 one of the possible configurations proposed by Robinson (2015). The tectonic evolution of North Qiangtang  
1255 and South Qiangtang terranes follows the in situ suture model (e.g. Zhang et al., 2006a,b, 2011); however,  
1256 the alternative underthrust model interpretation (e.g. Kapp et al., 2000, 2003a; Kapp, 2001) does not  
1257 significantly influence the paleogeographic reconstruction of western Tibet (see text for further  
1258 discussion). The dotted grey line in each diagram refers to the location of Fig. 10 for the same time  
1259 intervals. The black rectangle in (h) locates the study area in the framework of western Tibet, which is  
1260 enlarged in (i). (**i**) Proposed correlation between metamorphic and magmatic terranes of western Tibet and  
1261 those of Central Tibet and Pamir-Karakorum. The Bazar Dara Slates unit (BDS) is correlated to the Karakul  
1262 Mazar terrane, the Surukwat Complex (SC) and the Aghil Granodiorite (AG) body are correlated to the  
1263 South Qiangtang terrane, whereas the Shaksgam Sedimentary Belt (SSB) and the Sughet Granodiorite (SG)  
1264 body are correlated to the South Pamir terrane. Abbreviations for the other terranes, oceans and sutures  
1265 are as follows: KL: Kunlun; SG: Songpan-Ganze; KM: Karakul-Mazar; NQ: North Qiangtang; SQ: South  
1266 Qiangtang; A: Amdo; CP: Central Pamir; SP: South Pamir; KK: Karakorum; LH: Lhasa; KH/LK: Kohistan-  
1267 Ladakh; I: India; *KS*: Kunlun Suture; *JS*: Jinsha Suture; *TS*: Tanyamas Suture; *LSSZ*: Longmu Tso-Shuanghu  
1268 Suture Zone; *RPSZ*: Rushan-Pshart Suture Zone; *TBZ*: Tirich Boundary Zone; *BNSZ*: Bangong-Nunjiang Suture  
1269 Zone; *SSZ*: Shyok Suture Zone; *ITSZ*: Indus-Tsangpo Suture Zone; *KKF*: Karakorum Fault; *KF*: Karakax Fault;  
1270 *LCF/ATF*: Longmu Co Fault / Althyn Tagh Fault.

1 **Pre-Cenozoic evolution of the Aghil Range (western Tibetan Plateau): a missing piece of**  
2 **the Tibet-Pamir-Karakorum geopuzzle**

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23 **Abstract**

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24 The Tibetan Plateau, largely derived from the accretion of several Gondwana microplates to the southern  
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4 margin of Asia since the late Palaeozoic, is the highest and largest topographic relief on Earth. Although the  
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6 first order geodynamic processes responsible for its pre-Cenozoic evolution are quite well-known, many  
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8 issues are still debated, among which is the timing of collision of each terrane with the southern margin of  
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10 Asia. Even more uncertain is the pre-Palaeozoic history of these terranes, due to the lack of basement  
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12 exposures. As a contribution to understanding the pre-Cenozoic evolution of the Tibetan Plateau, this paper  
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14 focuses on the Aghil Range, a remote and poorly investigated area close to the Karakorum Fault between  
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16 Kunlun and Karakorum (Xinjiang, China) in western Tibet. The tectono-metamorphic and magmatic  
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18 evolution of the Aghil Range is investigated using a multidisciplinary approach that combines field mapping,  
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20 petrology and geochronology (U-Pb on titanite, zircon, monazite and xenotime using SHRIMP-RG). We  
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22 demonstrate that the Aghil Range preserves a coherent slice of Neoproterozoic crystalline basement with a  
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24 late Palaeozoic sedimentary cover deposited on a passive continental margin during the Gondwana break-  
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26 up. This represents the westernmost exposure of Precambrian crystalline basement known so far in the  
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28 Tibetan Plateau. Furthermore, petrological and geochronological results allow reconstructing the Mesozoic  
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30 poly-metamorphic evolution of this sector of the Tibetan Plateau, which records the evidence of Middle  
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32 Jurassic (ca. 170 Ma) and Late Cretaceous (66 Ma) collisional events, as well as of the Late Jurassic (ca. 150  
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34 Ma) early subduction of an accretionary complex developed on its southern margin. Evidence of Late  
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36 Cretaceous subduction-related magmatism preceding the last collisional event is also recorded. These  
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38 results allow tentative correlation of the different terranes of Central Tibet with those of the Pamir-  
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40 Karakorum Range on both sides of the Karakorum fault.

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41 **Key-words**

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43 Tibetan Plateau, pre-Cenozoic evolution, Cimmerian orogeny, petrology, U-Pb geochronology

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45 **Research highlights**

- 46 • The pre-Cenozoic evolution of the Aghil Range (western Tibet) is investigated
- 47 • A coherent slice of Neoproterozoic basement with its sedimentary cover is preserved
- 48 • A correlation between terranes of Central Tibet and Pamir-Karakorum is proposed

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56 **1. Introduction**

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58 The Tibetan Plateau, together with the Pamir-Karakorum Range, is the highest and largest topographic  
59 plateau on Earth: understanding its formation and evolution is therefore fundamental for clarifying the  
60 geodynamic processes leading to crustal thickening and continental growth. The origin, evolution and  
61 tectonic architecture of the Tibetan Plateau and of the Pamir-Karakorum Range have been studied for  
62 several decades (e.g. Sengör, 1979, 1987; Allégre et al., 1984; Sengör, 1987; Dewey et al., 1988; Gaetani et  
63 al., 1990a,b, 1993; Burtman and Molnar, 1993; Matte et al., 1996; Gaetani, 1997; Yin and Harrison, 2000);  
64 important advances in understanding their genesis have been made over the last ten years thanks to an  
65 increasing number of field, geochemical and geochronological studies as demonstrated by publication of  
66 numerous review papers (e.g. Zhang et al., 2012; Zhu et al., 2011, 2013) and special volumes dedicated to  
67 this topic (e.g. Zhang and Santosh, 2012; Zanchi et al., 2015; Chung and Niu, 2016; Zhang et al., 2017).

68 It is now widely accepted that the Tibetan Plateau and Pamir-Karakorum Range derive from the  
69 accretion of several Gondwana-derived microplates (also called Cimmerian terranes: Sengör, 1984) to the  
70 southern margin of Asia since the late Palaeozoic, in response to the ongoing subduction and progressive  
71 closure of the oceanic basins between each terrane (e.g. Allégre et al., 1984; Dewey et al., 1988; Yin and  
72 Harrison, 2000; Pan et al., 2012). Moreover, there is increasing evidence that most of the deformation,  
73 shortening and crustal thickening within the plateau were the result of these pre-Cenozoic accretionary  
74 processes whereas the contribution of India-Asia collision to the building of the plateau was only minor  
75 (e.g. Murphy et al., 1997; Yin and Harrison, 2000; Hildebrand et al., 2001; Robinson et al., 2004; Kapp et al.,  
76 2003b, 2005, 2007; Guynn et al., 2006; Zhang et al., 2012; Zhu et al., 2013). However, several recent studies  
77 explain crustal thickening as related to the underthrusting of Indian lithosphere (e.g. Chen et al., 2017).  
78 Although the first order geodynamic processes responsible for the pre-Cenozoic evolution of the plateau  
79 are quite well-known, many issues are still debated. These include the timing, duration and direction of  
80 oceanic subduction and the timing of collision of each terrane with the southern margin of Asia. Even more  
81 uncertain is the Precambrian history of the Cimmerian terranes, due in large part to the scarcity of  
82 basement exposures and the predominance of late Palaeozoic or younger supracrustal assemblages (e.g.  
83 Pan et al., 2004; Guynn et al., 2012; Zhu et al., 2013 and references therein).

84 Most studies devoted to understanding the pre-Cenozoic formation and evolution of the Tibetan  
85 Plateau and Pamir-Karakorum Range are based on either stratigraphic evidence (e.g. Gaetani et al., 1990b,  
86 1993; Gaetani, 1997; Kapp et al., 2007; Zanchi and Gaetani, 2011; Zanchi et al., 2012; Angiolini et al., 2013,  
87 2015; Gaetani and Leven, 2014; Robinson, 2015; Zeng et al., 2016) or on the distribution, composition and  
88 age of subduction- and/or collision- related magmatic rocks (e.g. Schwab et al., 2004; Zhu et al., 2011,  
89 2013; Zhang et al., 2012 and references therein; Zanchetta et al., 2018). Studies of metamorphic rocks are  
90 sporadic and limited to few areas, such as the Central Qiangtang Metamorphic Belt (e.g. Kapp et al., 2000,  
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91 2003a; Pullen et al., 2008, 2011; Liang et al., 2012; Zhai et al., 2011b; Zhao et al., 2014), the Amdo terrane  
92 (e.g. Guynn et al., 2006, 2012, 2013; Zhang et al., 2012a), the Central Lhasa terrane (e.g. Dong et al., 2011;  
93 Zhang et al., 2012b ; Zhang et al., 2014) and the southern Karakorum Terrane (e.g. Searle and Tirrull, 1991;  
94 Lemennicier et al., 1996; Fraser et al., 2001; Streule et al., 2009; Searle, 2011).

95 As a contribution to the understanding of the pre-Cenozoic evolution of the Tibetan Plateau, this  
96 study focuses on the metamorphic and magmatic units exposed along the Aghil Range, a remote and poorly  
97 investigated area located at the junction between the Tibetan Plateau and the Pamir-Karakorum Range in  
98 western Tibet. Interpretation of this area is still controversial, having been alternatively ascribed to the  
99 Songpan-Ganze (Tianshuai) terrane (e.g. Valli et al., 2008; Leloup et al., 2012; Pan et al., 2012) or to the  
100 Southern (or Western) Qiangtang terrane (e.g. Gaetani et al., 1990a; Robinson, 2009, 2015; Groppo and  
101 Rolfo, 2008; Streule et al., 2009; Searle et al., 2010, 2011; Gaetani and Leven, 2014; Rolfo et al., 2014). The  
102 aim of this study is therefore twofold: (i) to understand the geological significance of the Aghil Range and  
103 assess the nature of the boundaries between the various terranes on both sides of the Karakorum Fault; (ii)  
104 to tentatively reconstruct the pre-Cenozoic history of accretion, collision, metamorphism and magmatism  
105 of the western portion of the Tibetan Plateau, in comparison with the better known evolution of the central  
106 portion of the plateau.

107 We present field, petrographic, petrologic and geochronologic data on both metamorphic and  
108 magmatic rocks exposed along a geological transect, approximately 40 km long, located between the  
109 Kunlun to the north and the Karakorum to the south (Xinjiang, China). The studied metamorphic rocks  
110 belong to two different tectonic units (Gaetani et al., 1990a): the “Bazar Dara Slates”, a metasedimentary  
111 unit located few km south of Mazar (not to be confused with the thick terrigenous Bazardara Series of SE  
112 Pamir, which are tentatively correlated with the Singhiè Formation of the Shaksgam Sedimentary Belt by  
113 Gaetani and Leven, 2014), and the “Surukwat Complex”, a composite sequence of metamorphic thrust  
114 sheets derived from both magmatic and sedimentary protoliths. Two granodioritic bodies (i.e. Aghil  
115 Granodiorite and Sughet Granodiorite) tectonically interposed between these metamorphic units have  
116 been investigated as well.

117 The petrological and geochronological results allow us to: (i) clarify how the Cimmerian terranes are  
118 assembled on both sides of the Karakorum fault (i.e. in the western Tibetan Plateau and in the Pamir-  
119 Karakorum Range); (ii) demonstrate that the Surukwat Complex represents the westernmost exposure of a  
120 Precambrian crystalline basement known so far in the Tibetan Plateau; (iii) reconstruct the Mesozoic poly-  
121 metamorphic evolution of both the Bazar Dara Slates Unit and the Surukwat Complex which record the  
122 evidence of Middle Jurassic and Late Cretaceous collisional events, as well as the Late Jurassic early  
123 subduction of an accretionary complex developed on the southern margin of the Surukwat Complex; (iv)  
124 recognise evidence of Late Cretaceous subduction-related magmatism preceding the collision of the Lhasa  
125 and South Qiangtang terranes; and (v) confirm the diachronicity, from east to west, of the Cretaceous

126 collisional event between the Lhasa and South Qiangtang terranes in western Tibet and between the  
127 Kohistan-Ladakh and Karakorum terranes in Pamir-Karakorum .

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## 2. Geodynamic Setting of the Tibetan Plateau and Pamir-Karakorum Range

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### 2.1 Tibetan Plateau

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134 The Tibetan Plateau consists of four main E-W trending crustal terranes that rifted from the eastern margin  
135 of Gondwana during the late Palaeozoic, drifted northward across the Tethyan Ocean basins and then  
136 progressively accreted to the southern margin of Asia during the Mesozoic (e.g. Allégre et al., 1984; Sengör,  
137 1987; Yin and Harrison, 2000). In central Tibet, the following terranes are conventionally distinguished from  
138 north to south (Fig. 1a): Kunlun, Songpan-Ganze, Qiangtang and Lhasa terranes. The boundaries between  
139 these terranes coincide with different suture zones resulting from the closure of the ocean basins originally  
140 interposed between each terrane, and now marked by discontinuous belts of ophiolite fragments and  
141 m $\acute{e}$ lange:

- 142 – The Jinsha Suture (JS) separates the Songpan-Ganze terrane from the Qiangtang terrane and records  
143 closure of the Paleo-Tethys Ocean; the Songpan-Ganze terrane is commonly interpreted as an extensive  
144 arc-accretionary system built along the southern margin of Kunlun during the Triassic (e.g. Matte et al.,  
145 1996; Schwab et al., 2004), whose huge volume of sediments did not allow a complete continent-  
146 continent collision between Kunlun and Qiangtang (e.g. Roger et al., 2010).
- 147 – The Bangong-Nujiang Suture Zone (BNSZ), more than 1200 km long, separates the Qiangtang terrane  
148 from the Lhasa terrane and resulted from the closure of the Bangong-Nujiang Ocean (or Meso-Tethys  
149 Ocean; Sengör, 1984). In western and eastern Tibet, the BNSZ ophiolitic belt is doubled and isolates two  
150 micro-blocks interposed between Qiangtang and Lhasa terranes: the Risum block to the west and the  
151 Amdo terrane to the east. The Risum block is interpreted as an oceanic arc formed by the intra-oceanic  
152 subduction of the Bangong-Nujiang Ocean (Matte et al., 1996; Shi et al., 2004; Shi, 2007; Liu D. et al.,  
153 2017). The Amdo terrane is an old micro-continent within the Bangong-Nujiang Ocean that  
154 amalgamated with the Qiangtang block before the Lhasa–Qiangtang collision (Xu et al., 1985; Guynn et  
155 al., 2006). The Amdo block has been variably correlated to the Lhasa terrane (e.g. Coward et al., 1988;  
156 Harris et al., 1988; Yin and Harrison, 2000; Pan et al., 2004) or the Qiangtang terrane (e.g. Guynn et al.,  
157 2006, 2012; Zhu et al. 2013).
- 158 – The 2000-km-long Indus-Tsangpo Suture Zone (ITSZ) represents the site where the Neo-Tethys  
159 lithosphere separating the Lhasa terrane and north India was consumed at a subduction zone dipping  
160 northward beneath the Lhasa terrane (Yin and Harrison, 2000).

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160 – The Longmu Tso-Shuanghu Suture Zone (LSSZ) divides the Qiangtang terrane in two sub-terrane: the  
161 North (or Eastern) Qiangtang terrane and the South (or Western) Qiangtang terrane. This suture zone is  
162 spatially associated with a high-pressure metamorphic belt (Central Qiangtang Metamorphic Belt,  
163 CQMB: Kapp et al., 2000, 2003a), and its origin is still debated. The CQMB has been interpreted either:  
164 (i) as a part of the Songpan-Ganze accretionary mélange that was underthrust beneath a single  
165 Qiangtang terrane during the southward subduction of the Paleo-Tethys along the Jinsha Suture in the  
166 early Mesozoic (i.e. underthrust model: e.g. Kapp et al., 2000, 2003a; Kapp, 2001; Pullen et al., 2011), or  
167 (ii) as an *in situ* suture zone formed by northward subduction of a branch of Paleo-Tethys originally  
168 separating the South Qiangtang terrane of Gondwanan affinity from the North Qiangtang terrane of  
169 Cathaysian affinity (i.e. intra-Qiangtang suture model: e.g. Zhang, 2001; Zhang et al., 2006a,b, 2011;  
170 Zhang and Tang, 2009; Liu et al., 2011; Zhai et al., 2011a,b, 2013; Zhu et al., 2013; Zhao et al., 2014).

171 The timing, duration and direction of oceanic subduction, as well as the timing of final collision between  
172 each terrane are still debated. The Paleo-Tethys ocean was subducted northward beneath the Kunlun and  
173 Songpan-Ganze terranes by the Late Permian - Early Triassic, as evidenced by the occurrence of a magmatic  
174 belt and volcanic arc along the southern margin of Kunlun (e.g. Matte et al., 1996; Xiao et al., 2002, 2003;  
175 Liu et al., 2015; Cao et al., 2015); southward subduction in Late Triassic – Early Jurassic times is instead  
176 proposed by other studies (e.g. Kapp et al., 2000, 2003a; Zhang et al., 2016). The final collision between  
177 the Qiangtang and the Songpan-Ganze/Kunlun terranes occurred in the Late Triassic to Middle Jurassic. The  
178 southernmost Bangong-Nunjiang ocean was subducted either northward under the South Qiangtang  
179 terrane, or southward beneath the Lhasa Terrane during the Mesozoic (Zhu et al., 2013; Liu D. et al., 2017  
180 and references therein). This was likely an oblique subduction, resulting in a diachronous collision of the  
181 Lhasa and Qiangtang terranes from west (Middle Cretaceous) to east (Late Cretaceous) (e.g. Matte et al.,  
182 2006; Zhang et al., 2008; Zhu et al., 2013; Liu L. et al. 2017; Liu D. et al., 2017 and references therein).

## 2.2 Pamir-Karakorum Range

186 The dextral strike-slip Karakorum fault (KKF) (e.g. Phillips et al., 2004; Searle and Phillips, 2007; Phillips,  
187 2008; Valli et al., 2008; Robinson, 2009; Leloup et al., 2011) separates the Tibetan Plateau to the east from  
188 the Pamir-Karakorum Range to the west, whose tectonic framework is also the result of the amalgamation  
189 of different Gondwana-derived terranes. From north to south these terranes are the Kunlun, Karakul-Mazar  
190 (or North Pamir), Central Pamir, South Pamir, Karakoram and Kohistan-Ladakh, and they are bounded by  
191 the Kunlun, Tanyamas, Rushan-Pshart, Tirich Boundary Zone and Shyok suture zones, respectively (Fig. 1a).  
192 The correlation between crustal terranes and suture zones of Pamir-Karakorum and central Tibet is not  
193 univocal (e.g. Burtman and Molnar, 1993; Yin and Harrison, 2000; Lacassin et al., 2004; Schwab et al., 2004;  
194 Robinson, 2009, 2015; Robinson et al., 2012; Zanchetta et al., 2018). The Karakul-Mazar terrane, bounded

195 by the Kunlun suture to the north and the Tanyamas Suture to the south, is commonly considered the  
196 equivalent of the Songpan-Ganze terrane of central Tibet. Interpretation of the other terranes is less  
197 certain. Recent studies suggest that the Central Pamir, South Pamir and Karakoram terranes are correlative  
198 to the Qiangtang terrane (e.g. Phillips et al., 2004; Upadhyay et al., 2005; Searle and Phillips, 2007;  
199 Robinson et al., 2004; Robinson, 2009, 2015; Searle et al., 2010; Searle, 2011; Angiolini et al., 2013; Yang et  
200 al., 2017) and that the Kohistan-Ladakh terrane and the Lhasa terrane of central Tibet are part of a  
201 continuous magmatic arc, built on a continental basement in the Lhasa terrane and in the Ladakh region,  
202 and on oceanic crust in the Kohistan terrane (e.g. Rolfo et al., 1997; Yin and Harrison, 2000; Robinson et al.,  
203 2004; Robinson, 2009, 2015). Following this interpretation, the Shyok Suture, that separates the Karakorum  
204 terrane from the Kohistan-Ladakh terrane, would be the equivalent of the Bangong-Nunjiang suture zone.  
205 However, other studies suggest that South Pamir is the equivalent of the Lhasa terrane (e.g. Lacassin et al.,  
206 2004; Schwab et al., 2004), in which case the Rushan-Pshart Suture Zone would be equivalent to the  
207 Bangong-Nunjiang suture zone. An absence of direct correlations between major suture zones east and  
208 west of the Karakorum Fault has been also recently proposed (Zanchetta et al., 2018). A better  
209 understanding of the western Tibet region and the Aghil Range where the study area is located in  
210 particular, is crucial for meaningful correlation of crustal terranes across the Karakorum fault.

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### 3. Geological Setting

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215 The Aghil Range in western Tibet (Fig. 1a) was studied along a cross-section between Ilik (at the confluence  
216 between the Yarkand and the Zug Shaksgam rivers) and Sughet Jagal (northern K2 base camp, along the  
217 Sarpo Laggo Valley), across the Aghil pass and the Shaksgam Valley (Fig. 1b). Very few geological studies  
218 have been performed in this area, tectonically sandwiched between Kunlun to the north and Karakorum to  
219 the south. The first “modern”, though preliminary, data are reported in Gaetani et al. (1990a, 1991); these  
220 results already documented the complexity of the area, characterized by the juxtaposition of different  
221 metamorphic, sedimentary or magmatic units. More recently, Groppo and Rolfo (2008) reported the  
222 evidence of a possibly old metamorphic basement north of Aghil Pass (i.e. the Surukwat Complex), and  
223 constrained the P-T evolution of its structurally upper portion. However, the nature and age of this  
224 basement is still unknown, as well as the age of metamorphism. The non-metamorphic Shaksgam  
225 Sedimentary Belt was investigated in detail by Gaetani et al. (1990a, 1991) and Gaetani and Leven (2014),  
226 whereas the magmatic rocks exposed close to the Aghil Pass (i.e. the Aghil Granodiorite) were studied very  
227 recently by Liu L. et al. (2017). Detailed description of the lithological and tectonic architecture along the Ilik  
228 – Sughet Jagal transect and relevant images of field geology and mesoscopic structures are given by Rolfo  
229 et al. (2014) and will only be summarized here.

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229 Starting from Ilik and going upstream (southward) along the Aghil Dara Valley, the Bazar Dara Slates  
230 Unit (Fig. 1b) is a metasedimentary sequence consisting mainly of phyllites and metasandstones that are  
231 dipping steeply towards SSE and locally intruded by undeformed Late Triassic granitic bodies (e.g. Mazar  
232 Granodiorite; Liu et al., 2015). A sub-vertical to south-dipping fault separates the Bazar Dara Slates Unit  
233 from the Surukwat Complex; this is a composite sequence of thrust sheets steeply dipping SSW (Fig. 1b,d)  
234 in which, except for a few non-metamorphic slivers, there is a general southward increase in metamorphic  
235 grade from lower to higher structural levels (Rolfo et al., 2014). Although pervasively mylonitized, the  
236 internal coherence of this basement is mostly preserved (Fig. 1c). From north to south, the Surukwat  
237 Complex starts with a non-metamorphic sliver of red sandstone with anhydrite interlayers that shows a  
238 strong affinity with the Qiangtang terrane (Leeder et al., 1988), and is petrographically similar to the  
239 Jurassic Marpo Sandstone of the Shaksgam Valley (Gaetani et al., 1990a, 1991). A metamorphic basement  
240 derived from igneous protoliths is thrust over the red sandstone to the south (Fig. 1d). From lower to  
241 higher structural levels it consists of meta-diorite, meta-granodiorite and meta-granite with transposed  
242 meta-aplitic dykes. Primary intrusive relationships between the various magmatic protoliths have been  
243 obliterated by the pervasive mylonitic deformation which affected this portion of the Surukwat Complex. A  
244 small (few-meters-thick) slice of slightly metamorphosed limestone is tectonically intercalated in this meta-  
245 igneous basement. A few-kilometers-thick layer of metaconglomerate dominated by granitic and dioritic  
246 clasts follows further upstream. Further south, the metaconglomerate is overlain by a km-thick  
247 metasedimentary sequence that consists of quartzite, quartzitic gneiss and metapelite with meter-thick  
248 intercalations of meta-marl (amphibole-bearing calcschist, biotitic-amphibolitic schist, carbonate-rich  
249 garnet-bearing biotitic schist) and impure marble. The medium grade metapelites represent the last unit of  
250 the Surukwat Complex, which is bounded to the south by a sharp fault contact with the Aghil intrusive body  
251 (Fig. 1b,d). A weakly deformed apophysis of this pluton crops out within the quartzitic gneisses to the  
252 north. The Aghil intrusive body mainly consists of amphibole + biotite –bearing granodiorite, with minor  
253 monzonite and porphyritic granite, the latter cropping out in the proximity of the Aghil Pass. A Late  
254 Cretaceous age has been recently obtained for these magmatic rocks (zircon U-Pb ages; Liu L. et al., 2017).

255 Another tectonic contact is crossed southward just before the Aghil Pass (Fig. 1b,d) that separates  
256 the Aghil Granodiorite from the Shaksgam Sedimentary Belt (Gaetani et al., 1990a, 1991), a ~250 km long  
257 and 15-20 km wide sedimentary sequence exposed along the Shaksgam Valley and extending to the west of  
258 the Karakorum fault for ~150 km in southeast Pamir (Robinson, 2009). This sedimentary sequence,  
259 displaced in a system of open folds, faulted, thrust and stacked together, is at least 3 km thick and spans  
260 from the Lower Permian to Jurassic (Gaetani et al., 1990a, 1991; Gaetani, 1997; Gaetani and Leven, 2014).  
261 The main strand of the Karakorum Fault is crossed a few kilometres south of the junction between the  
262 Sarpo Laggo and Shaksgam valleys (Fig. 1b,d). A cataclastic contact separates the Shaksgam Sedimentary  
263 Belt (Aghil Dolomite) from the Sughet Granodiorite. This plutonic body, mostly made of biotite-bearing

264 granodiorite, crops out in the lower (northern) part of the Sarpo Laggo valley, near Sughet Jagal, and  
265 represents the last unit of the investigated geological transect.

266 This study focuses on six representative samples from the Bazar Dara Slates Unit (BDS: sample 06-  
267 01), the Surukwat Complex (SC: samples 06-10, 06-17 and 06-115) and the Aghil and Sughet magmatic  
268 bodies (samples 06-26 and 06-108). These samples have been selected out of a total of 76 samples (BDS; 16  
269 samples; SC: 54 samples; magmatic bodies: 6 samples) after careful petrographic characterization of the  
270 entire sample suite.

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## 4. Methods

### 4.1 Mineral chemistry

Minerals were analysed with a Cambridge Stereoscan 360 SEM equipped with an EDS Energy 200 and a Pentafet detector (Oxford Instruments) at the Department of Earth Sciences, University of Torino. The operating conditions were as follows: accelerating voltage was set to 15 kV, beam diameter was 2  $\mu\text{m}$ , and detection limits for oxides were 0.03 wt%. SEM–EDS quantitative data were acquired and processed using the Microanalysis Suite Issue 12, INCA Suite version 4.01; natural mineral standards were used to calibrate the raw data; the  $\rho\phi Z$  correction (Pouchou and Pichoir, 1988) was applied. Absolute error is 1  $\sigma$  for all calculated oxides. Mineral chemical data of representative minerals are reported in Table 1.

### 4.2 Determination of peak P-T conditions in the metamorphic samples

P-T conditions for all the metamorphic samples were estimated using the “Average PT” routine of THERMOCALC (Holland and Powell, 1998, version 3.33, thermodynamic database 5.5). Activity-composition relationships were calculated using AX. This method, which estimates the optimal P-T conditions using a set of independent reactions that fully describe the thermodynamics of the system, is able to find a result only if the given mineral assemblage defines a sufficient number of reactions between end-members. The obtained results were considered reliable if passed the ‘sigfit’ test ( $\sigma_{\text{fit}} < \text{cutoff value}$ ), giving P–T uncertainties ( $\sigma T$  and  $\sigma P$ ) at  $\pm 1\sigma$  (95% confidence). End member(s) with erratic behaviour (large  $e^*$  values) and a low influence on the least squares results (low  $hat$  values) were removed from the calculation because they may cause inconsistency in the results (see Powell and Holland, 1994). A pure  $\text{H}_2\text{O}$  fluid was considered in the calculations. Average P-T results are reported in Table 2.

The pseudosection approach cannot be applied on most of the studied samples. Sample 06-1 (Bazar Dara Slates phyllite) contains significant amounts of calcite, mostly concentrated in late veins, and it is not

299 possible to obtain the equilibrium bulk composition needed for the pseudosection calculation. Similarly, an  
300 equilibrium bulk composition cannot be determined for samples 06-10 and 06-17 (meta-granodiorite and  
301 meta-diorite from the Surukwat Complex), which are clearly not completely re-equilibrated, as suggested  
302 by the presence of relict magmatic hornblende crystals (see section 4.1.2). Therefore, the pseudosection  
303 approach has been applied only on sample 06-115 (two-micas, garnet-bearing micaschist from the  
304 Surukwat Complex). The pseudosection for this sample was already calculated by Groppo and Rolfo (2008)  
305 and it was used to constrain its complete P-T evolution.

### 307 **4.3 Geochronology and Trace Element Analyses**

309 U/Pb geochronology was performed on zircon separated from igneous samples (06-26, 06-108) and  
310 titanite, monazite, and xenotime in-situ on polished thin sections of metamorphic rocks (06-1, 06-10, 06-17,  
311 06-115), using the SHRIMP-RG (sensitive high-resolution ion microprobe with reverse geometry) instrument  
312 at the Stanford–U.S. Geological Survey Micro-Analysis Center at Stanford University.

313 Heavy mineral separates from granitoid samples 06-26 and 06-108 were obtained by standard  
314 pulverizing, magnetic and heavy liquid methods. Individual zircon grains were handpicked under alcohol,  
315 mounted in epoxy resin with natural zircon standards and polished to expose the grain centers for analysis  
316 by secondary ion microprobe spectrometry (SIMS). Zircon grains were imaged by cathodoluminescence (CL)  
317 to expose intra-grain zoning or complexity and aid in placing SIMS spots. The U-Pb and trace element  
318 analysis (Tables 3 and SM1a) was performed simultaneously following routines outlined in Barth and  
319 Wooden (2006, 2010). Instrument mass fractionation corrections were calibrated by replicate analysis of  
320 the zircon standard R33 zircon (419 Ma; Black et al., 2004) with a  $2\sigma$  calibration error for the R33  $^{206}\text{Pb}/^{238}\text{U}$   
321 ratio of 0.69% for the analytical session added in quadrature. Ages were calculated from  $^{206}\text{Pb}/^{238}\text{U}$  ratios  
322 corrected for common Pb using the  $^{207}\text{Pb}$  method using measured  $^{207}\text{Pb}/^{206}\text{Pb}$  ratios or using the  $^{204}\text{Pb}$   
323 method (see Williams, 1998). The U concentration was calibrated with Madagascar Green (MAD-559; 3940  
324 ppm U, Coble et al., 2018). Data reduction and plotting utilized programs Squid 2.51 and Isoplot 3.76 of  
325 Ludwig (2009, 2012).

326 The acquisition routine included  $^{89}\text{Y}$ ,  $^{139}\text{La}$ ,  $^{140}\text{Ce}$ ,  $^{146}\text{Nd}$ ,  $^{147}\text{Sm}$ ,  $^{153}\text{Eu}$ ,  $^{155}\text{Gd}$ ,  $^{163}\text{Dy}^{16}\text{O}$ ,  $^{166}\text{Er}^{16}\text{O}$ ,  
327  $^{172}\text{Yb}^{16}\text{O}$ ,  $^{90}\text{Zr}_2^{16}\text{O}$ , and  $^{180}\text{Hf}^{16}\text{O}$  simultaneous with U/Pb analysis. In a separate analytical session, additional  
328 trace element analyses were performed, including  $^{27}\text{Al}$ ,  $^{30}\text{Si}$ ,  $^{31}\text{P}$ ,  $^{39}\text{K}$ ,  $^{40}\text{Ca}$ ,  $^{28}\text{Si}^{16}\text{O}$ ,  $^{45}\text{Sc}$ ,  $^{48}\text{Ti}$ ,  $^{49}\text{Ti}$ ,  $^{56}\text{Fe}$ ,  $^{89}\text{Y}$ ,  
329  $^{93}\text{Nb}$ ,  $^{94}\text{Zr}^{16}\text{O}$ , and  $^{96}\text{Zr}$ . Each isotope was normalized to  $^{28}\text{Si}^{16}\text{O}$  or  $^{90}\text{Zr}_2^{16}\text{O}$ , and concentrations were  
330 calibrated against zircon standard MAD-559 (Coble et al., 2018). The estimated errors based on repeated  
331 analysis of MAD-559 was 6 to 10 % for P, Sc, Ti, Y and Nb. The uncertainty of Al, K, Ca, and Fe were higher  
332 (up to 45% RSD), but these elements were measured only to evaluate if the analytical spot intersected an  
333 inclusion or alteration. For example, grain 0608-2.1, Sc, Nb, and Ti were omitted because Al, K and Fe were

334 ~20 to 100x higher than other zircon from the same sample. Chondrite normalized plots were calculated  
335 using values from McDonough and Sun (1995). The  $^{49}\text{Ti}$  data were used to determine the Ti content to  
336 avoid interference of  $^{96}\text{Zr}^{2+}$  with the  $^{48}\text{Ti}$  peak (Watson and Harrison, 2005). Ti-in-zircon temperatures were  
337 calculated using Ferry and Watson (2007), assuming the activity of  $\text{SiO}_2$  is equal to one ( $a_{\text{SiO}_2} = 1$ ) and  
338 activity of  $\text{TiO}_2$  is approximately 0.7 ( $a_{\text{TiO}_2} = 0.7$ ) for rutile-absent siliceous melts (Hayden and Watson,  
339 2007).

340 Titanite (samples 06-10 and 06-17), monazite (06-115, 06-1), and xenotime (06-115, 06-1) were  
341 analyzed *in-situ* in polished thin sections that were cut into fragments and mounted with appropriate  
342 natural standards in large format epoxy mounts (megamounts). Elemental maps showing U, Y, Ce, P, Ca,  
343 and Th concentrations of monazite and high contrast backscatter electron (BSE) images of titanite,  
344 monazite and xenotime were generated to identify zoning prior to analysis. Monazite element maps were  
345 collected on a JEOL JXA-0823 Electron Microprobe at the University of Iowa and BSE imaging was  
346 performed at Stanford University using a JEOL 5600 SEM.

347 U-Pb analysis of titanite (Tables 4 and SM1b) monazite (Tables 5 and SM1c), and xenotime (Tables 6  
348 and SM1d) followed the same analytical routine used for zircon, except  $^{89}\text{Y}$  was not included in the  
349 acquisition table for xenotime and only U and Th were measured as trace elements for titanite. U-Pb ages  
350 were standardized relative to 44069 monazite (425 Ma; Aleinikoff et al., 2006), MG-1 xenotime (490 Ma;  
351 Fletcher et al., 2010; Aleinikoff et al., 2012), and MMs titanite (524 Ma; Schoene and Bowring, 2006)  
352 reference materials for monazite, xenotime, and titanite unknown samples, respectively. For trace  
353 elements, each isotope was normalized  $^{140}\text{Ce}^{31}\text{P}^{16}\text{O}_2$ ,  $^{89}\text{Y}^{16}\text{O}_2$ , or  $^{40}\text{Ca}^{48}\text{Ca}^{48}\text{Ti}^{16}\text{O}_2$ , and trace element  
354 concentrations were standardized relative to 44069 monazite (calibrated relative to Namibia (NAM)  
355 monazite; Aleinikoff et al., 2012), MG-1 xenotime, or BLR titanite (Aleinikoff et al., 2007) for monazite,  
356 xenotime, and titanite, respectively.

357 Common Pb composition for titanite samples was determined by linear regression of all analyses on  
358 a 3D Tera-Wasserburg plot which yielded a data-defined  $^{207}\text{Pb}/^{206}\text{Pb}$  upper intercept of 0.9096, interpreted  
359 as the best estimate of the common Pb composition. All other mineral use initial common Pb isotopic  
360 composition approximated from Stacey and Kramers (1975). Data reduction for geochronology follows the  
361 methods described by Williams (1998) and Ireland and Williams (2003), using the MS Excel add-in programs  
362 Squid2.51 and Isoplot3.76 of Ken Ludwig (2009, 2012). For titanite, the  $^{206}\text{U}/^{238}\text{U}$  calibration constant  
363 utilized a data-defined slope of 1.24 through the distribution of MMs analyses on a plot of  $\ln(\text{UO}^+/\text{U}^+)$  vs  
364  $\ln(\text{Pb}^+/\text{U}^+)$ . Zircon was calculated with a fixed slope of 2.0. Monazite and xenotime analysis used energy  
365 filtering to eliminate the isobaric interference at mass  $^{204}\text{Pb}$ , and a calibration of  $\ln(\text{UO}_2^+/\text{UO}^+)$  vs  
366  $\ln(\text{Pb}^+/\text{UO}^+)$  with a data-defined slope (0.60 and 2.14, respectively), following methods modified from  
367 Fletcher et al. (2010) and Cross and Williams (2018).

## 369 5. Results

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### 371 5.1 Petrography and petrology of the studied samples

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#### 373 5.1.1 Bazar Dara Slates Unit (sample 06-1)

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The Bazar Dara Slates Unit exposed along the lower Aghil Dara Valley consists of a metasedimentary sequence of sandstones, siltstones and slates, steeply dipping towards SSE and locally rich in deformed quartz + carbonate veins. Sample 06-1, collected ca. 4 km south-east of Ilik (N36°22'28.4" E76°40'54.4" – 3560 m), is a two-micas + chlorite phyllite (meta-sandstone) consisting of quartz, muscovite, biotite, chlorite, calcite, minor albite and accessory ilmenite (Fig. 2a). Most calcite is a late phase, concentrated in millimetric veins either concordant or discordant with respect to the main foliation; however, it cannot be excluded that minor calcite is also present in the equilibrium assemblage. The main foliation ( $S_m$ ), defined by the preferred orientation of muscovite ( $Si = 3.13-3.25$  a.p.f.u. on the basis of 11 oxygens), chlorite ( $X_{Mg} = 0.53-0.54$ ), biotite ( $X_{Mg} = 0.51-0.54$ ;  $Ti = 0.11-0.12$  a.p.f.u.) and ilmenite, is pervasively crenulated, with the local appearance of an  $S_{m+1}$  defined by white mica and ilmenite. Minor monazite and xenotime occur as anhedral grains with no clear relationship relative to the dominant foliation (Fig. 3).

Equilibrium P-T conditions for this sample were calculated for the  $S_m$  assemblage  $Qz + Ab + Chl + Mu + Bt + Ilm$ , which resulted in  $320 \pm 32$  °C,  $5.2 \pm 0.9$  kbar (i.e. greenschist-facies conditions) (Table 2). The rare occurrence of relict phengite flakes ( $Si = 3.32-3.39$  a.p.f.u.) partially replaced by muscovite suggests a complex metamorphic evolution, possibly characterized by an earlier high-pressure stage (pre- $S_m$ :  $Qz + Ab + Chl + Phe$ ).

#### 391 5.1.2 Surukwat Complex (samples 06-10, 06-17, 06-115)

##### 392 4.1.2.1 Meta-diorite and meta-granodiorite (samples 06-10 and 06-17)

The lowermost portion of the Surukwat Complex consists of a sequence of strongly mylonitized metabasites of dioritic to granodioritic composition with sub-vertical attitude, alternating with granitic to aplitic layers (Fig. 1d). Sample 06-10 and 06-17 are representative examples of the most and less deformed lithologies, respectively.

Sample 06-10 (N36°17'54,8" E76°35'22,8" – 3830 m) is a mylonitized hornblende-bearing meta-granodiorite still preserving mineralogical relics of the igneous protolith (Fig. 2b). It is characterized by mm-sized, sharply zoned amphibole porphyroclasts (Fig. 2b,c), with a yellow-pale green relic core (i.e.  $Amp_1$ : magmatic Mg-hornblende to edenite;  $Si = 6.6-7.1$  a.p.f.u.,  $Al^{IV} = 1.1-1.4$  a.p.f.u.,  $Al^{VI} = 0.2-0.5$  a.p.f.u.,  $X_{Mg} = 0.55-0.70$ ) surrounded by a light green rim ( $Amp_2$ : metamorphic actinolite;  $Si = 7.5-7.8$ ,  $Al^{IV} = 0.2-0.6$  a.p.f.u.,  $Al^{VI} = 0.0-0.15$  a.p.f.u.,  $X_{Mg} = 0.65-0.80$ ). A very thin outermost rim of Mg-hornblende ( $Amp_3$ :  $Si = 7.2-7.4$  a.p.f.u.,  $Al^{IV} = 0.6-0.8$  a.p.f.u.,  $Al^{VI} = 0.15-0.32$  a.p.f.u.,  $X_{Mg} = 0.65-0.72$ ) is locally observed. Amphibole

404 porphyroclasts are wrapped around by the pervasive mylonitic foliation (Fig. 2b,c), mainly defined by  
405 phengitic white mica (Si = 3.30-3.37 a.p.f.u.) locally rimmed by muscovite (Si = 3.20-3.30 a.p.f.u.),  
406 associated with chlorite ( $X_{Mg} = 0.65-0.67$ ), quartz, albite ( $Ab_{96-100}$ ) and epidote ( $Ps_{20-26}$ ). Titanite, allanite and  
407 minor rutile occur as accessory minerals. Titanite is present both as large (100-200  $\mu m$  in size) subhedral  
408 grains with common opaque intergrowths, interpreted as relics of the magmatic protolith, and as small (<  
409 10  $\mu m$ ) euhedral grains aligned in the main foliation, interpreted as metamorphic (Fig. 4a). Due to the very  
410 small size of the metamorphic titanite, only the larger magmatic grains have been dated.

411 Sample 06-17 (N36°17'33.0" E76°34'58.2" - 3870m) is a poorly deformed, hornblende + biotite-  
412 bearing meta-diorite, still preserving the porphyric structure of the protolith (Fig. 2d). The deep-green, mm-  
413 sized, magmatic hornblende ( $Amp_1$ : magmatic Fe-hornblende; Si = 6.5-7.3 a.p.f.u.,  $Al^{IV} = 0.7-1.5$  a.p.f.u.,  $Al^{VI}$   
414 = 0.3-0.6 a.p.f.u.,  $X_{Mg} = 0.34-0.50$ ) is surrounded by a light green actinolitic rim ( $Amp_2$ : metamorphic  
415 actinolite; Si = 7.5-7.8,  $Al^{IV} = 0.2-0.4$  a.p.f.u.,  $Al^{VI} = 0.1-0.3$  a.p.f.u.,  $X_{Mg} = 0.54-0.56$ ), in turn overgrown by a  
416 thin outermost rim of deep-green Fe-hornblende ( $Amp_3$ : metamorphic Fe-hornblende; Si = 6.8-7.1 a.p.f.u.,  
417  $Al^{IV} = 0.9-1.2$  a.p.f.u.,  $Al^{VI} = 0.4-0.6$  a.p.f.u.,  $X_{Mg} = 0.38-0.45$ ) (Fig. 2e,f). The former plagioclase phenocrysts of  
418 the protolith are replaced by mm-sized, slightly zoned albite porphyroblasts (core:  $Ab_{97-100}$ ; rim:  $Ab_{95-97}$ )  $\pm$   
419 epidote ( $Ps_{25-30}$ ), whereas magmatic biotite is replaced by fine-grained aggregates of greenish-brown biotite  
420 ( $X_{Mg} = 0.40-0.42$ ; Ti = 0.12-0.13 a.p.f.u.) + epidote ( $Ps_{15-20}$ ) + minor chlorite ( $X_{Mg} = 0.45-0.47$ ) (Fig. 2f).  
421 Amphibole and albite porphyroblasts are set in a fine-grained matrix consisting of epidote + albite + minor  
422 quartz. Titanite occurs as large (up to 1 mm) aggregates of fine-grained euhedral crystals (Fig. 4b-d),  
423 interpreted as relics of the magmatic protolith.

424 The observed mineral assemblages and compositions suggest that both samples 06-10 and 06-17  
425 preserve the evidence of two distinct metamorphic events. The first event was more pervasive and was  
426 responsible for the growth of the actinolitic rim at the expenses of the magmatic hornblende in both  
427 samples, in equilibrium with albite + epidote + chlorite,  $\pm$  phengite (sample 06-10),  $\pm$  biotite (sample 06-17);  
428 the second event is marked by the appearance of the hornblende outermost rim in equilibrium with albite  
429 + epidote,  $\pm$  muscovite and chlorite in sample 06-10, and  $\pm$  biotite in sample 06-17. Equilibrium P-T  
430 conditions of the first metamorphic event were calculated for the assemblages Act + Ab + Phe + Chl + Ep +  
431 Qtz + Rt + Ttn of sample 06-10, which resulted to be  $482 \pm 20$  °C,  $11.5 \pm 1.2$  kbar (i.e. transition between  
432 greenschist- and blueschist-facies conditions; Fig. SM1); the second metamorphic event was constrained at  
433  $512 \pm 30$  °C,  $4.5 \pm 1.7$  kbar (i.e. transition between greenschist- and amphibolite-facies conditions; Fig. SM1)  
434 using the Hbl + Ab + Mu + Chl + Ep + Qtz + Rt + Ttn assemblage of sample 06-10 (Table 2). Mineral  
435 assemblages of sample 06-17 do not define enough reactions for Average PT to work, but are nevertheless  
436 consistent with the results obtained from sample 06-10. Overall, these data suggest a clockwise P-T  
437 evolution characterized by relatively high-P peak conditions of  $\sim 480$ °C, 11 kbar followed by decompression  
438 coupled with moderate heating at  $\sim 510$ °C, 4.5 kbar.

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### 5.1.2.2 Two-micas, garnet-bearing, metapelite (sample 06-115)

At its uppermost structural level, the Surukwat Complex mostly consists of metapelitic lithologies with minor intercalations of biotite-rich amphibolites and impure marbles likely derived from marl and limestone protoliths, respectively (Fig. 1d). Two of these metapelites, among which sample 06-115 studied in this work, have been petrologically investigated by Groppo and Rolfo (2008). Sample 06-115 is a two-micas, garnet-bearing micaschist consisting of quartz, muscovite, biotite, garnet, plagioclase ( $An_{5-18}$ ), chlorite and accessory ilmenite (Fig. 2g). The main schistosity ( $S_m$ ) is defined by the alignment of muscovite ( $Si = 3.00 - 3.10$  a.p.f.u.), biotite ( $X_{Mg} = 0.42-0.50$ ;  $Ti = 0.09-0.11$  a.p.f.u.) and ilmenite, and derives from the transposition of an earlier foliation ( $S_{m-1}$ ) defined by the same phases and still preserved in the microlithons (Fig. 2g). Garnet porphyroblasts ( $Alm_{71-75}Sp_{12-15}Prp_{7-9}Grs_{2-3}$ ) are enveloped by the  $S_m$  and overgrow the  $S_{m-1}$ , still preserved as an internal foliation (Fig. 2g). The outermost garnet rim is characterized by a sharp increase in  $X_{Mn}$  ( $Alm_{69-71}Sp_{20-21}Prp_{6-7}Grs_{2-3}$ ), likely reflecting diffusional re-equilibration at the onset of the  $S_m$  development. Late chlorite flakes ( $X_{Mg} = 0.45-0.47$ ) statically overgrow the  $S_m$ .

Monazite and rare xenotime are present as accessory minerals. Monazite occurs as subhedral grains within garnet and as elongate grains and clusters of grains aligned parallel to the dominant foliation in the matrix (Fig. 5).

The results of thermodynamic modeling (pseudosection approach: Groppo and Rolfo, 2008) suggest that the peak assemblage (Grt + Wm + Bt + Pl + Qz + Ilm) grew at 580-600 °C, 8-9 kbar; consistent results are given by the Average PT method applied on the same assemblage ( $645 \pm 26$  °C,  $8.2 \pm 1.2$  kbar) (Table 2; Fig. SM1). Basing on the pseudosection results, Groppo and Rolfo, 2008 further constrained the main foliation development at ca. 500 °C, 5 kbar. Furthermore, combining the results obtained from sample 06-115 with those obtained from a second metapelite sample (06-117), the same authors inferred a steep and narrow anticlockwise P-T evolution for this portion of the Surukwat Complex (Fig. SM1).

### 5.1.3 Aghil and Sughet magmatic bodies (samples 06-26, 06-108)

Sample 06-26 (N36°11'02.4" E76°37'32.7" - 4905 m) was collected from the Aghil Granodiorite body, in the proximity of the Aghil Pass. It is a biotite-bearing porphyritic granite crosscutting the main granodiorite, with perthitic K-feldspar crystals up to several centimeters in length (Fig. 2h). Plagioclase is zoned, with the Ca-richer core locally altered in sericite  $\pm$  saussurite, and a thin albite rim. Biotite is partially replaced by chlorite. Zircon and apatite are abundant among the accessory minerals and are often included in biotite.

Sample 06-108 (N36°04'08.3" E76°24'52.4" - 3887 m) was collected from the Sughet Granodiorite body, near Sughet Jagal. It is a biotite-bearing granodiorite with poikilitic K-feldspar, zoned plagioclase partially altered in sericite, and brown biotite pervasively replaced by chlorite (Fig. 2i); apatite, zircon, and allanite occur as accessory minerals.

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## 5.2 Geochronology results

### 5.2.1 Zircon

Zircon from the Aghil (06-26) and Sughet (06-108) granodiorite bodies is euhedral with well-developed oscillatory zoning (Fig. 6). Sample 06-26 yielded reproducible ages and consistent REE patterns (Fig. 7a). Eight analyses define a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $83 \pm 1$  Ma (MSWD=1.6) (Fig. 7a and Table 3). Results from sample 06-108 exhibit more scatter with one Proterozoic and 11 Cretaceous apparent ages. The Proterozoic age is interpreted as a xenocrystic core. Seven of the younger ages define a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $102 \pm 1$  Ma (MSWD=0.9; Fig. 7b, Table 3). Two slightly older analyses are interpreted to reflect mixture of xenocrystic cores and younger magmatic zircon. The two younger analyses are inferred to record younger disturbance possibly associated with emplacement of the latest magmatic products of the Sughet granodiorite suite.

### 5.2.2 Titanite

Titanite  $^{206}\text{Pb}/^{238}\text{U}$  analyses from samples 06-10 and 06-17 (Surukwat Complex) generally contain low U (average 2 ppm U) and high common Pb resulting in mixing array on Tera-Wasserburg plots (Fig. 8). Regression of analyses from sample 06-17 yields a well-defined 3-D isochron intercept age of  $796 \pm 29$  Ma (MSWD=1.7), that is consistent with a weighted mean  $^{206}\text{Pb}/^{238}\text{U}$  age of  $772 \pm 31$  Ma (MSWD=2.4) for the same data. All analyses from sample 06-10 (large titanite grains) are dominated by common Pb, plotting near the upper intercept, with a regression that defines a lower intercept of  $635 \pm 410$  Ma, consistent with results of sample 06-17. Titanite ages from both samples are interpreted to record the protolith age of the orthogneiss bodies.

### 5.2.3 Xenotime and monazite

Xenotime analyses from sample 06-1 (Bazar Dara Slates Unit) define a weighted mean age of  $174 \pm 11$  Ma (MSWD = 2.8) (Fig. 9a, Table 6) and show a middle REE (MREE) enriched patterns (Fig. 9a) with strong depletion in light REE (LREE) characteristic of metamorphic xenotime (e.g. Aleinikoff et al., 2015). The xenotime grains occur in the matrix and have an ambiguous relationship with the dominant foliation. The xenotime ages are interpreted to record metamorphism at ca. 170 Ma. Two matrix monazite grains in sample 06-1 yield  $^{206}\text{Pb}/^{238}\text{U}$  ages of  $153 \pm 3$  Ma and  $161 \pm 4$  Ma, which we interpret to be consistent with the ca. 170 Ma metamorphic age derived from xenotime, whereas a much younger metamorphic age of  $35 \pm 17$  Ma is recorded by a single matrix monazite.

In sample 06-115 (Surukwat Complex), xenotime is very rare. One grain of xenotime gave a  $^{204}\text{Pb}$  corrected  $^{206}\text{Pb}/^{238}\text{U}$  age of  $165 \pm 6$  Ma, which is consistent with the ca. 170 Ma metamorphic age derived



509 from xenotime in sample 06-1. Monazite in sample 06-115 dominantly occurs in the matrix but is locally  
510 preserved within garnet (Fig. 5). The monazite shows negative Eu anomalies and heavy REE depletion (Fig.  
511 9b) consistent with growth in the presence of feldspar and garnet, respectively (e.g. Rubatto et al., 2013).  
512 Two monazite  $^{206}\text{Pb}/^{238}\text{U}$  age populations occur in this sample: (i) two monazite grains included in garnet  
513 gave  $^{206}\text{Pb}/^{238}\text{U}$  ages of  $144 \pm 4$  Ma and  $157 \pm 6$  Ma (Fig. 9b), thus defining a ca. 150 Ma monazite  
514 generation; (ii) nine analyses from matrix monazite in sample 06-115 (Fig. 9b, Table 5) define a weighted  
515 mean age of  $66 \pm 2$  Ma (MSWD = 1.1) (second monazite generation). The remaining grains record ages  
516 between 108 and 70 Ma that are interpreted to reflect partial resetting of the older monazite formed  
517 originally at ca. 150 Ma. This interpretation is supported by core to rim Th zoning observed in grains that  
518 give U/Pb ages >70 Ma (Fig. 5).

## 6. Discussion

### 6.1 Pre-Cenozoic geodynamic evolution of the Aghil Range

#### 6.1.1 The Surukwat Complex: a relic of Neoproterozoic basement with a possible late Palaeozoic cover

525 The U-Pb titanite ages obtained for the meta-granodiorite and meta-diorite (samples 06-10 and 06-17)  
526 from the Surukwat Complex tightly constrain the formation of the igneous protoliths at  $796 \pm 29$  Ma, i.e.  
527 during the Neoproterozoic.

528 Although not common, fragments of Neoproterozoic crystalline basement have been reported from  
529 different terranes in the central Tibetan Plateau (see Zhu et al., 2013 for a review). In the central Lhasa and  
530 Amdo terranes, U-Pb dating of zircons in gneissic rocks yielded 787-748 Ma (Nam Tso area: Hu et al., 2005)  
531 and 920-820 Ma (Amdo: Guynn et al., 2006, 2012) ages, respectively. Pre-Cambrian basement exposures  
532 are not reported so far from the Qiangtang terrane, the oldest basement rocks having been dated as  
533 Ordovician (Pullen et al., 2011; Zhao et al., 2014) or Cambro-Ordovician (Kapp et al., 2000). However, the  
534 existence of a > 740 Ma old basement in the Qiangtang terrane is suggested by Neoproterozoic inherited  
535 zircon ages obtained from an Early-Cretaceous granite from the Longmu Co area, western Tibet (Leloup et  
536 al., 2012). Neoproterozoic ages (ca. 865-825 Ma) have been also obtained for fragments of crystalline  
537 basement exposed in the southern part of the Songpan-Ganze terrane, in eastern Tibet (Roger and  
538 Calassou, 1997; Zhou et al., 2002, 2006a,b). Relicts of a possibly pre-Cambrian basement are also reported  
539 from the Karakorum (e.g. Le Fort et al., 1994; Zanchi and Gaetani, 2011), South Pamir (East Hindu Kush)  
540 (e.g. Zanchi and Gaetani, 2011) and Central Pamir (e.g. Schwab et al., 2004) terranes of the Pamir-  
541 Karakorum Range. All these data suggest that the different Gondwana-derived terranes which constitute  
542 the Tibetan Plateau and the Pamir-Karakorum Range are characterized by similar pre-Cambrian igneous  
543 basements, likely formed during the initial stages of Rodinia break-up (e.g. Guynn et al., 2012 and

544 references therein). Our data allow the recognition of a new exposure of such Pre-Cambrian basement in  
545 the Surukwat Complex: this finding represents the westernmost occurrence, known so far, of Pre-Cambrian  
546 basement rocks in the Tibetan Plateau.

547 The Neoproterozoic basement of the Surukwat Complex is overlaid by a meta-sedimentary cover  
548 whose protoliths (i.e. conglomerate immediately overlying the igneous basement followed upward by  
549 sandstone and then mudstone with marl and limestone intercalations) are compatible with deposition in a  
550 passive continental margin setting (Fig. 1c). The age of these sedimentary protoliths is unknown, but it  
551 should be not younger than Triassic, because our data demonstrate that during Middle Jurassic these  
552 sediments were already metamorphosed (see below, Section 6.1.2). Basing on the strong similarities with  
553 late Palaeozoic meta-sedimentary sequences overlying the crystalline basement in the central Lhasa, Amdo  
554 and Qiangtang terranes (Kapp et al., 2000, 2003b, 2007; Leier et al., 2007; Pullen et al., 2008; Guynn et al.,  
555 2012), we suggest that the protoliths of the meta-sedimentary sequence exposed in the upper portion of  
556 the Surukwat Complex might be late Palaeozoic in age. In the absence of geochronological data  
557 constraining the age of the sedimentary protoliths, however, this hypothesis remains speculative, but is  
558 nevertheless consistent with the observed lithological associations. Our results thus suggest that the  
559 Surukwat Complex is a coherent slice of Neoproterozoic crystalline basement of dioritic to granodioritic and  
560 granitic composition, with a possibly late Palaeozoic sedimentary cover (Fig. 1c) deposited on a passive  
561 continental margin during the Gondwana break-up.

### 562 563 *6.1.2 The Middle Jurassic collision between the Bazar Dara Slates and the Surukwat Complex and the* 564 *Cimmerian Orogeny*

565 The first tectono-metamorphic event experienced by the studied units is Middle Jurassic (ca. 170 Ma). This  
566 event is registered by xenotime and monazite in the two micas phyllite of the Bazar Dara Slates Unit  
567 (sample 06-1), and by rare xenotime grains in the garnet-bearing micaschist from the Surukwat Complex  
568 Unit (sample 06-115).

569 Evidence of a Late Triassic - Middle Jurassic orogenic event are widespread in both the Tibetan  
570 Plateau and the Pamir-Karakorum Range; this event, known as Cimmerian Orogeny (Sengör, 1984), led to  
571 the final amalgamation of the Cimmerian terranes (i.e. Qiangtang, Amdo, Central Pamir, South Pamir and  
572 Karakorum terranes) to the southern margin of Asia (i.e. Songpan-Ganze/Karakul Mazar and Kunlun  
573 terranes) through the complete closure of the Paleo-Tethys, and of the smaller Longmu Tso-Shuanghu,  
574 Rushan-Pshart and Wakan oceanic basins. The timing of each collision has been mostly determined basing  
575 on either stratigraphic or magmatic constraints, whereas metamorphic constraints have been less  
576 frequently used. Previous studies have shown that the final collision between each terrane occurred at  
577 slightly different ages from west to east. In the central Tibetan Plateau, the final collision of the Qiangtang  
578 terrane with the Songpan-Ganze and Kunlun terranes occurred during the Late Triassic - Early Jurassic (e.g.  
579

579 Dewey et al., 1988; Matte et al., 1996; Roger et al., 2003, 2010; Dai et al., 2013; Cao et al., 2015; Liu et al.,  
580 2015), and the amalgamation of South Qiangtang and North Qiangtang was broadly coeval, independently  
581 from the model chosen to explain the nature of the Central Qiangtang Metamorphic Belt (i.e. underthrust  
582 model vs intra-Qiangtang suture model, see section 2). In the Pamir-Karakorum Range the accretions of  
583 Central Pamir to Karakul Mazar (North Pamir) and of South Pamir to Central Pamir were broadly coeval at  
584 ca. 200 Ma, i.e. during the latest Triassic - earliest Jurassic (e.g. Gaetani et al., 1993; Zanchi et al., 2000,  
585 2012; Zanchi and Gaetani, 2011; Angiolini et al., 2013; Robinson et al., 2015; Zanchetta et al., 2018),  
586 whereas the collision of Karakoram with South Pamir is supposed to be Late Early Jurassic (e.g. Searle and  
587 Tirrull, 1991; Gaetani et al., 1993; Angiolini et al., 2013).

588 In this framework, we therefore interpret the Middle Jurassic xenotime and monazite ages  
589 recorded by the two micas phyllite of the Bazar Dara Slates Unit (sample 06-1) to be related to the  
590 Cimmerian orogenic event, i.e. to the collision between the Bazar Dara Slates and the Surukwat Complex  
591 (Fig. 10); according to the petrologic results, this collisional event was responsible for a greenschist-facies  
592 (i.e. ~320°C, 5 kbar) metamorphic overprint in the Bazar Dara Slates Unit (Fig. 11a,b). This is consistent with  
593 the stability field of both xenotime and monazite, that can grow in low-grade pelitic schists (Pan, 1997;  
594 Wing et al., 2003; Bollinger and Janots, 2006; Janots et al., 2006; Rasmussen et al., 2007; Krenn and Finger,  
595 2007).

596 Due to the scarcity of xenotime in the metapelite of the Surukwat Complex (sample 06-115), the  
597 interpretation of its age and of the P-T conditions of its growth remain uncertain. It has been demonstrated  
598 that xenotime remains stable with increasing metamorphic grade in garnet-absent rocks, whereas it reacts  
599 out continuously as garnet grows, and is typically gone from the matrix assemblage in samples from the  
600 middle garnet zone (Spear and Pyle, 2002; Fitsimons et al., 2005). This constrains the growth of xenotime at  
601 temperatures below the first appearance of garnet (i.e.  $T < 500^{\circ}\text{C}$  according to the pseudosection modelling  
602 in Groppo and Rolfo, 2008) and explains why xenotime is so scarce in the garnet-bearing sample 06-115.  
603 The ca. 170 Ma age obtained from the rare xenotime grains in this sample might be thus related to the  
604 same Cimmerian orogenic event registered by the Bazar Dara Slates Unit. An alternative hypothesis could  
605 be to correlate the ca. 170 Ma age to the earliest stage of development of an accretionary complex on the  
606 southern margin of the Surukwat Complex (see below, section 5.1.3; Fig. 11a).

### 6.1.3 *The Late Jurassic formation of an accretionary prism on the southern margin of the Surukwat Complex*

609 The Surukwat Complex experienced another significant tectono-metamorphic event during the Late  
610 Jurassic at ca. 150 Ma, responsible for the growth of a first generation of monazite in the metapelite  
611 sample 06-115 (two grains:  $157 \pm 6$  Ma and  $144 \pm 4$  Ma). Microstructural observations show that this  
612 monazite generation is preserved in garnet, whereas in the matrix it is partially reset by the growth of a  
613 younger monazite generation (Fig. 5). It has been demonstrated that the first appearance of monazite in

614 metapelitic rocks is strongly influenced by the bulk rock composition, particularly by the CaO content (e.g.  
615 Foster and Parrish, 2003; Wing et al., 2003) and the CaO/Na<sub>2</sub>O ratio (e.g. Janots et al., 2008), and to a lesser  
616 extent by the Al<sub>2</sub>O<sub>3</sub> content and FeO/MgO ratio (e.g. Fitsimons et al., 2005). Janots et al. (2008)  
617 demonstrated that in metapelites with low Ca/Na ratios (i.e. CaO/Na<sub>2</sub>O<0.54), monazite is formed through  
618 the complete breakdown of allanite at T > 580°C, whereas higher Ca/Na ratios enhance the allanite stability  
619 towards higher temperatures. The CaO/Na<sub>2</sub>O ratio of sample 06-115 is 0.26 (Groppo and Rolfo, 2008),  
620 therefore it is likely that in this sample monazite grew at T > 580°C, through a reaction such as that  
621 proposed by Janots et al. (2008) (i.e. allanite + apatite + Al-Fe-Mg phases<sub>1</sub> → monazite + anorthite + Al-Fe-  
622 Mg phases<sub>2</sub>, with chlorite and garnet involved as Al-Fe-Mg phase<sub>1</sub> and phase<sub>2</sub>, respectively). Garnet  
623 growth was modelled at ~600 °C, 9 kbar (i.e. at peak conditions; Groppo and Rolfo, 2008), therefore we  
624 suggest that the ca. 150 Ma registered by monazite grains included in garnet might be associated to the  
625 peak metamorphic event, and that monazite formation was nearly coeval with garnet growth (Fig. 11d), as  
626 also evidenced by its trace element pattern (Fig. 9b).

627 To interpret the geodynamic significance of this Late Jurassic metamorphic event, the apparently  
628 contrasting anticlockwise and clockwise P-T evolutions inferred for samples 06-115 and 06-10/17 should be  
629 considered (Groppo and Rolfo, 2008). Anticlockwise P-T paths at relatively high P-T conditions are  
630 characteristic of accretionary systems during the early stages of underflow, as suggested for the Franciscan  
631 melange (e.g. Platt, 1975; Cloos, 1985, 1986; Ernst, 1988; Krogh et al., 1994), the Coastal Cordillera of  
632 south-central Chile (e.g. Willner et al., 2004; Willner, 2005; Hyppolito et al., 2014) and, possibly, the North  
633 America Cordillera (Perchuck et al., 1999). In such geodynamic setting, rocks that were piled up earlier to  
634 the hot hanging wall of the subduction channel in the deepest part of the accretionary complex, were  
635 metamorphosed at higher temperatures for a given pressure (typically at albite-epidote amphibolite-facies  
636 conditions, transitional between high-pressure greenschist facies and eclogite facies; ~600-650 °C, 8-12  
637 kbar), compared to material subducted later (Perchuck et al., 1999). In the following accretionary stages,  
638 the hanging wall became progressively cooler and isotherms were continuously displaced to greater depth  
639 due to the continuous de-hydration of the later subducted material, thus implying that the early subducted  
640 rocks were exhumed at temperatures lower than those experienced during subduction, whereas the rocks  
641 involved later in the accretionary process recorded a normal clockwise P-T path (e.g. Willner et al., 2004;  
642 Willner, 2005; Hyppolito et al., 2014). Anticlockwise P-T metamorphic evolutions related to the inception of  
643 subduction are not commonly observed because of the extremely small areas affected by this type of  
644 metamorphism (≤ tens of km<sup>2</sup>, with thickness of few hundred meters: Wakabayashi, 2004). P-T conditions  
645 and evolutions recorded by the studied samples fit well with this geodynamic scenario (Fig. 10e-f, 11c-f).  
646 Moreover, it has been recently demonstrated that in similar tectonic settings, the accretionary process did  
647 not result in a chaotic mixing, but rather in the formation of a “pseudo”-coherent unit (Hyppolito et al.,

648 2014). This is also consistent with our observation that the Surukwat Complex consists of a tectonic  
649 intermingling of lithological units which nevertheless maintain their internal coherence.

650 To summarize, we suggest that during the Late Jurassic – Early/Middle Cretaceous, an accretionary  
651 prism developed on the southern margin of the Surukwat Complex (Fig. 10e-f). The first material involved in  
652 the subduction was a portion of the upper Palaeozoic (?) sedimentary cover (i.e. sample 06-115), which was  
653 subducted in a relatively hot environment at ca. 150 Ma (Fig. 11c,d). The successive subduction of the  
654 thinned southern portion of the pre-Cambrian basement (i.e. samples 06-10 and 06-17) occurred in a  
655 cooler environment, during the early exhumation of the firstly subducted metasedimentary rocks (Fig.  
656 11e,f).

#### 658 *6.1.4 The Late Cretaceous collision between the Surukwat Complex and the Lhasa terrane*

659 The Surukwat Complex registered the last tectono-metamorphic event during the Late Cretaceous: this  
660 event is recorded in the metapelite sample 06-115 by the growth of monazite at  $66 \pm 2$  Ma. This second  
661 generation of monazite is ubiquitous in the matrix and appears in equilibrium with the main foliation,  
662 whose P-T conditions of formation were constrained by Groppo and Rolfo (2008) at  $\sim 500^\circ\text{C}$ , 5 kbar.

663 Evidence of Late Cretaceous metamorphism has been reported from western Tibet and Pamir-  
664 Karakorum Range and are interpreted as the result of collision between the Lhasa and South Qiangtang  
665 terranes (western Tibet) and between the Kohistan-Ladakh and Karakorum terranes (Pamir-Karakorum),  
666 respectively. More specifically, in western Tibet, tight constrains on the time of this collision have been  
667 recently presented by Liu L. et al. (2017) and Liu D. et al. (2017) basing on the Late Cretaceous age of syn-  
668 orogenic magmatism. In the Pamir-Karakorum Range, the timing of collision between Kohistan-Ladakh and  
669 Karakourm has been constrained as Middle to Late Cretaceous basing on stratigraphic data (e.g. Gaetani et  
670 al., 1990b, 1993; Robinson et al., 2004; Zanchi and Gaetani, 2011; Gaetani, 2016), age of syn-orogenic  
671 magmatism (e.g. Fraser et al., 2001; Heuberger et al., 2007; Searle and Philipps, 2007; Searle, 2011) and  
672 metamorphic studies (e.g. Fraser et al., 2001; Streule et al., 2009; Searle, 2011). A complex and long-lasting  
673 history of tectonic and magmatic activity associated with the Kohistan-Karakorum suture zone, possibly  
674 extending up to Eocene time, is also proposed by Heuberger et al. (2007).

675 The  $66 \pm 2$  Ma monazite age recorded by sample 06-115 can be therefore interpreted as the  
676 evidence of this collisional event (Fig. 10g, 11g-h). In this framework, the  $83 \pm 1$  Ma zircon age obtained  
677 from the porphyric granite of the Aghil Granodiorite body (sample 06-26) is perfectly compatible with a  
678 subduction-related syn-orogenic magmatism preceding the final collision between the Surukwat Complex  
679 and the Lhasa terrane (Fig. 11g). It is worth noting that Liu L. et al. (2017) obtained very similar results (78-  
680 80 Ma) from granitic rocks collected from the same area. A similar interpretation can be extended also to  
681 the  $102 \pm 1$  Ma zircon age obtained from the Sughet Granodiorite (sample 06-108): however, the Sughet

682 Granodiorite intrusive body is located south of the Karakorum fault and it is therefore not directly  
683 correlated to the geodynamic history of the Surukwat Complex (see below, section 5.2).

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685

## 685 **6.2 Solving the Tibet-Pamir-Karakorum geopuzzle**

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687

687 The new field, petrological and geochronological data presented in this paper provide new insights toward  
688 a better understanding of the Tibet-Pamir-Karakorum geopuzzle, allowing to correlate the different  
689 terranes of Central Tibet with those of the Pamir-Karakorum Range on both sides of the Karakorum fault.

690 We propose that (Fig. 12):

691 (1) the Bazar Dara Slates Unit might be the equivalent of the Sonpan-Ganze terrane of central Tibet, as  
692 already suggested by other authors (e.g. Gaetani et al., 1990a, 1991; Matte et al., 1996; Gaetani, 1997;  
693 Schwab et al., 2004; Rolfo et al., 2014) and can be correlated to the Karakul-Mazar terrane of the  
694 Pamir-Karakorum Range. Lithological and geochronological data from the Karakul-Mazar terrane of the  
695 eastern portion of Northern Pamir are especially similar to those described here for the Bazar Dara  
696 Slates Unit. These similarities include: (i) the occurrence of Triassic granitoids intruded in  
697 metasedimentary lithologies in both terranes (Robinson et al. 2007; Liu et al., 2015); (ii) the age of  
698 metamorphism (ranging from greenschist- to amphibolite-facies conditions) was constrained to be  
699 Early to Middle Jurassic (160-200 Ma) in the schists exposed along the Muztaghata massif of eastern  
700 Northern Pamir (Robinson et al., 2012) and is remarkably similar to the Middle Jurassic age of 170 Ma  
701 registered by the studied sample 06-1, which also experienced greenschist-facies metamorphism.  
702 Robinson et al. (2012) interpreted this age as the timing of the collision between the Karakul-Mazar  
703 terrane and the Central Pamir during final closure of the Paleo-Tethys Ocean, as well as we interpret  
704 the Middle Jurassic age as dating the collision between the Bazar Dara Slates and the Surukwat  
705 Complex.

706 (2) the Surukwat Complex might be correlated to the southern margin of the South Qiangtang terrane of  
707 central Tibet (see also Gaetani et al., 1990a, 1991; Rolfo et al., 2014), and shows significant similarities  
708 with the tectono-stratigraphy and metamorphic evolution of the Amdo terrane. These similarities  
709 include: (i) the occurrence of a Neoproterozoic basement in both the Amdo and Surukwat Complex. In  
710 both cases this basement consists of felsic to intermediate gneisses derived from igneous protoliths  
711 and of a metasedimentary cover derived from late Palaeozoic sediments deposited in a passive margin  
712 setting (Guynn et al., 2006, 2012); (ii) widespread evidence of Jurassic metamorphism preceding the  
713 Lhasa-Qiangtang collision. Peak metamorphic conditions in the Amdo terrane vary from ~600°C, 8 kbar  
714 (remarkably similar to peak conditions estimated for sample 06-115) to ~700°C, 10 kbar, and  
715 metamorphic peak has been dated as Early Jurassic (ca. 178 Ma: Guynn et al., 2006, 2013). The Amdo  
716 terrane thus experienced peak metamorphism earlier than the Surukwat Complex (ca. 150 Ma),

717 consistently with a diachronous activity of the Bangong-Nunjiang suture zone from east to west (Fig.  
718 12d-f).

719 Opposite to what preliminary suggested in Groppo and Rolfo (2008), the Surukwat Complex cannot be  
720 correlated to the Central Qiangtang Metamorphic Belt (CQMB), for three main reasons: (i) the CQMB  
721 mostly consists of a tectonic melange (e.g. Kapp et al., 2000, 2003a), whereas the Surukwat Complex is  
722 a coherent portion of an old basement with its original sedimentary cover still preserved; (ii) peak  
723 metamorphic conditions in some blocks of the CQMB reached blueschist and eclogite facies conditions  
724 (e.g. Kapp et al., 2000, 2003a; Zhang et al., 2006a,b, 2011; Pullen et al., 2008, 2011; Zhang and Tang,  
725 2009; Zhai et al., 2011b; Zhao et al., 2014), thus reflecting a significantly different geothermal gradient  
726 with respect to the Surukwat Complex; (iii) the timing of metamorphism in the CQMB is Middle to Late  
727 Triassic (244-223 Ma: Pullen et al., 2008, 2011) whereas in the Surukwat Complex it is Late Jurassic (ca.  
728 150 Ma).

729 (3) From a paleogeographic point of view, the Surukwat Complex occupies an intermediate position  
730 between the South Qiangtang terrane and the South Pamir terrane (e.g. Robinson, 2015, Chapman et  
731 al., 2018 and references therein), preserving different margins of the same micro-plate. The  
732 fundamental difference between the Surukwat Complex and South Pamir is that the latter does not  
733 show the evidence of Cretaceous metamorphism, having been “protected” from the collision with the  
734 Kohistan-Ladakh terrane by the interposed Karakorum terrane (Fig. 12g). Instead, the Karakorum  
735 terrane registered an important metamorphic and deformational event during the Late Cretaceous  
736 (between 83 and 62 Ma), that was interpreted as due to the collision with the Kohistan-Ladakh Arc  
737 (Fraser et al., 2001).

738 (4) In this framework, we propose that the non-metamorphic Shaksgam Sedimentary Belt and the Sughet  
739 Granodiorite body, now tectonically interposed between the Surukwat Complex and the Karakorum  
740 terrane, do not belong neither to the Surukwat Complex nor to the Karakorum terrane, but instead  
741 they are part of the South Pamir terrane (see also Robinson, 2009 and Gaetani and Leven, 2014). In  
742 other words, their actual location would not reflect their original position in pre-Cenozoic times, but it  
743 would be related to the offset effects of the Karakorum fault. This hypothesis is supported by both  
744 petrologic and geochronologic data showing that the Shaksgam Sedimentary Belt escaped the Late  
745 Cretaceous metamorphic and deformational event experienced by the Surukwat Complex and likely  
746 related to the collision with the Lhasa Terrane (Fig. 10g, 12g). Moreover, the  $102 \pm 1$  Ma zircon age  
747 obtained from the Sughet Granodiorite body is perfectly compatible with the Late Cretaceous  
748 calcalkaline subduction-related magmatism responsible for the emplacement of large intrusive bodies  
749 in South Pamir as well as in the North Karakorum terrane (e.g. Karakorum Batholith, 106-95 Ma)  
750 (Debon et al., 1987; Fraser et al., 2001; Schwab et al., 2004; Zanchi and Gaetani, 2011), which has been  
751 related to a north-directed low-dipping subduction below Karakorum.

752 **Supplementary material**

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753 Supplementary data to this article including (i) phase diagrams supporting the P-T evolution constrained for  
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754 the studied samples (Fig. SM1) and (ii) the complete set of U-Pb geochronologic data for zircon (Table  
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755 SM1a), titanite (Table SM1b), monazite (Table SM1c) and xenotime (Table SM1d) can be found online at:  
6  
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10  
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1168 **FIGURE CAPTIONS**

1169  
 1170 **Fig. 1 - (a)** Simplified tectonic map of the central-western Tibetan Plateau and Pamir-Karakorum Range  
 1171 (modified after Gaetani et al., 1991 and Robinson, 2009), with location of the Aghil Range (black rectangle).  
 1172 From north to south, the main terranes and sutures of the Tibetan Plateau are TB: Tarim Basin; KL: Kunlun;  
 1173 SG: Songpan-Ganze; NQ: North Qiangtang; SQ: South Qiangtang; LH: Lhasa; I: India; KS: Kunlun Suture; JS:  
 1174 Jinsha Suture; LSSZ: Longmu Tso-Shuanghu Suture Zone; BNSZ: Bangong-Nunjiang Suture Zone; ITSZ: Indus-  
 1175 Tsangpo Suture Zone, whereas those of the Pamir- Karakorum Range are KM: Karakul-Mazar (North Pamir);  
 1176 CP: Central Pamir; SP: South Pamir; KK: Karakorum; KH/LK: Kohistan-Ladakh; TS: Tanyamas Suture; RPSZ:  
 1177 Rushan-Pshart Suture Zone; TBZ: Tirich Boundary Zone; SSZ: Shyok Suture Zone. KKF: Karakorum Fault; KF:  
 1178 Karakax Fault; LCF/ATF: Longmu Co Fault / Althyn Tagh Fault. **(b)** Geologic map of the Aghil Range, between  
 1179 Kunlun (Yarkand River) and Karakorum (Shaksgam River), modified after Gaetani et al. (1991) and Rolfo et  
 1180 al. (2014), with location of the studied samples (black stars). **(c)** Sketch of the pre-Mesozoic original  
 1181 relationships between the crystalline basement and the sedimentary cover of the Surukwat Complex. **(d)**  
 1182 Geologic cross-section from Ilik to Sughet Jagal (A-B in (b)), with location of the studied samples (white  
 1183 stars).

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 1185 **Fig. 2 – Representative microstructures of the studied samples. Bazar Dara Slates Unit - Sample 06-1: (a)**  
 1186 Two-micas + chlorite phyllite: the main foliation  $S_m$ , defined by white mica + chlorite + biotite + ilmenite, is  
 1187 locally pervasively crenulated with the appearance of an  $S_{m+1}$  defined by white mica + ilmenite. Plane  
 1188 Polarized Light (PPL). *Surukwat Complex – Sample 06-10: (b)* The mylonitic foliation, defined by phengite,  
 1189 wraps around pluri-mm amphibole porphyroclasts (PPL). **(c)** Detail of an amphibole porphyroclast: the  
 1190 brownish-green amphibole core ( $Amp_1$ ) is a relic of the igneous protolith and it is rimmed by a green  
 1191 metamorphic actinolite ( $Amp_2$ ) (PPL). *Sample 07-17: (d)* The porphyric structure of the dioritic protolith is  
 1192 still preserved (note the mm-sized amphibole porphyroblasts set in a fine-grained matrix of albite + epidote  
 1193 + biotite ± chlorite) (PPL). **(e, f)** Details of strongly zoned amphibole crystals, with a dark green core ( $Amp_1$ ,  
 1194 magmatic hornblende), a light green rim ( $Amp_2$ , metamorphic actinolite) and a discontinuous deep green  
 1195 outermost rim ( $Amp_3$ , metamorphic hornblende) (PPL). *Sample 06-115: (g)* The main foliation  $S_m$ , defined  
 1196 by white mica + biotite, wraps around mm-sized garnet porphyroblasts. An earlier  $S_{m-1}$  is preserved in the  
 1197 microlithons and as an internal foliation within garnet. Chlorite porphyroblasts overgrows the  $S_m$  (PPL).  
 1198 *Plutonic bodies – Sample 06-26: (h)* This porphyritic granite is characterized by cm-sized perthitic K-  
 1199 feldspar; biotite is fresh and plagioclase is slightly zoned (Crossed Polarized Light: XPL). *Sample 06-108: (i)* In  
 1200 this biotite-bearing granodiorite, K-feldspar is poikilitic and biotite is mostly replaced by chlorite (XPL).

1202 **Fig. 3** – BSE images of monazite **(a, b)** and xenotime **(c, d)** in sample 06-1 showing location of U/Pb-TE  
1203 analysis. Scale bar in all images is 50  $\mu\text{m}$ .

1204  
1205 **Fig. 4** – BSE images of titanite from samples 06-10 **(a)** and 06-17 **(b-d)** showing location of U/Pb analysis.

1206  
1207 **Fig. 5** – BSE images **(a-c)** and compositional maps **(d)** of monazite from sample 06-115 showing location of  
1208 U/Pb-TE analysis. Monazite grains in (a) are included within garnet and yield Jurassic U/Pb ages. Matrix  
1209 monazite grains (b-c) are Late Cretaceous in age. Matrix monazite that displays core to rim zoning in Th (see  
1210 4-2.2 and 6-1.1 in d) gives older U/Pb ages interpreted to reflect incomplete recrystallization of Jurassic  
1211 monazite.

1212  
1213 **Fig. 6** – Cathodoluminescence (CL) images of zircons from samples 06-108 **(a)** and 06-26 **(b)** showing  
1214 location of U/Pb-TE analysis (large circle) and additional TE analysis (small circle).

1215  
1216 **Fig. 7** – Tera-Wasserburg plots of U/Pb data and plots of chondrite normalized trace element data from  
1217 zircon in samples 06-26 **(a)** and 06-108 **(b)**. Black ellipses and black symbols in chondrite normalized plots  
1218 are used in age interpretation; gray ellipses are not. Ellipses plotted at  $1\sigma$ . Age uncertainties reported at  
1219 95% confidence level (MSWD = mean square of weighted deviates). Tera-Wasserburg diagrams were made  
1220 using Isoplot (Ludwig, 2003). Chondrite normalized plots were calculated using values from McDonough  
1221 and Sun (1995).

1222  
1223 **Fig. 8** – Tera-Wasserburg plots of U/Pb data from titanite in samples 06-17 **(a)** and 06-10 **(b)**. Ellipses are  
1224 plotted at  $1\sigma$ . Lower intercept age uncertainties reported at 95% confidence level (MSWD = mean square of  
1225 weighted deviates). Tera-Wasserburg diagrams were made using Isoplot (Ludwig, 2003).

1226  
1227 **Fig. 9** – Tera-Wasserburg plots of U/Pb data and plots of chondrite normalized trace element data from  
1228 xenotime, sample 06-1 **(a)** and monazite, sample 06-115 **(b)**. Black ellipses and black symbols in chondrite  
1229 normalized plots are used in age interpretation; gray ellipses are not. Ellipses plotted at  $1\sigma$ . Analysis 5-2.1  
1230 is not plotted in (b) due large uncertainty in the  $^{207}\text{Pb}/^{206}\text{Pb}$  ratio. Age uncertainties reported at 95%  
1231 confidence level (MSWD = mean square of weighted deviates). Tera-Wasserburg diagrams were made  
1232 using Isoplot (Ludwig, 2003). Chondrite normalized plots were calculated using values from McDonough  
1233 and Sun (1995).

1234  
1235 **Fig. 10** – Pre-Cenozoic tectonic evolution of Western Tibetan Plateau as inferred from geochronological and  
1236 petrological data discussed in this paper. The sketch is especially focused on the evolution of the Surukwat

1237 Complex, whereas it is simplified for the other terranes. Abbreviations for the western Tibet terranes and  
1238 for the interposed oceans are: KL: Kunlun; BDs: Bazar Dara Slates Unit; SC: Surukwat Complex; LH: Lhasa; I:  
1239 India; *PT*: Paleo-Tethys; *BN*: Bangong-Nunjian Ocean; *NT*: Neo-Tethys. Abbreviations for suture zones are:  
1240 *KS*: Kunlun Suture; *JS*: Jinsha Suture; *BNSZ*: Bangong-Nunjian Suture Zone. The black rectangle refers to Fig.  
1241 11.

1242  
1243 **Fig. 11** – Enlargement of Fig. 10 focusing on the Mesozoic evolution of the Surukwat Complex and  
1244 explaining the development of the accretionary wedge on its southern margin (**a, c, e, g**). Colour codes are  
1245 the same as in Fig. 10. The P-T diagrams in (**b, d, f, h**) show the peak P-T conditions experienced by the  
1246 Bazar Dara Slates (b) and by the different thrust sheets of the Surukwat Complex (d, f, h) as inferred from  
1247 the Average PT results (with 1 $\sigma$  error ellipses). The P-T path of sample 06-115 is derived from Groppo and  
1248 Rolfo (2008), whereas that for sample 06-10 is tentatively inferred basing on the Average PT results.

1249  
1250 **Fig. 12** – (**a-h**) Simplified paleogeographic sketches showing the relative positions of the Cimmerian  
1251 Terranes in pre-Cenozoic times, as inferred from the data presented in this paper. The separation between  
1252 Central Pamir, South Pamir and Karakorum terranes in the Pamir-Karakorum Range is in agreement with  
1253 one of the possible configurations proposed by Robinson (2015). The tectonic evolution of North Qiangtang  
1254 and South Qiangtang terranes follows the in situ suture model (e.g. Zhang et al., 2006a,b, 2011); however,  
1255 the alternative underthrust model interpretation (e.g. Kapp et al., 2000, 2003a; Kapp, 2001) does not  
1256 significantly influence the paleogeographic reconstruction of western Tibet (see text for further  
1257 discussion). The dotted grey line in each diagram refers to the location of Fig. 10 for the same time  
1258 intervals. The black rectangle in (h) locates the study area in the framework of western Tibet, which is  
1259 enlarged in (i). (**i**) Proposed correlation between metamorphic and magmatic terranes of western Tibet and  
1260 those of Central Tibet and Pamir-Karakorum. The Bazar Dara Slates unit (BDS) is correlated to the Karakul  
1261 Mazar terrane, the Surukwat Complex (SC) and the Aghil Granodiorite (AG) body are correlated to the  
1262 South Qiangtang terrane, whereas the Shaksgam Sedimentary Belt (SSB) and the Sughet Granodiorite (SG)  
1263 body are correlated to the South Pamir terrane. Abbreviations for the other terranes, oceans and sutures  
1264 are as follows: KL: Kunlun; SG: Songpan-Ganze; KM: Karakul-Mazar; NQ: North Qiangtang; SQ: South  
1265 Qiangtang; A: Amdo; CP: Central Pamir; SP: South Pamir; KK: Karakorum; LH: Lhasa; KH/LK: Kohistan-  
1266 Ladakh; I: India; *KS*: Kunlun Suture; *JS*: Jinsha Suture; *TS*: Tanyamas Suture; *LSSZ*: Longmu Tso-Shuanghu  
1267 Suture Zone; *RPSZ*: Rushan-Pshart Suture Zone; *TBZ*: Tirich Boundary Zone; *BNSZ*: Bangong-Nunjiang Suture  
1268 Zone; *SSZ*: Shyok Suture Zone; *ITSZ*: Indus-Tsangpo Suture Zone; *KKF*: Karakorum Fault; *KF*: Karakax Fault;  
1269 *LCF/ATF*: Longmu Co Fault / Althyn Tagh Fault.



Figure 1  
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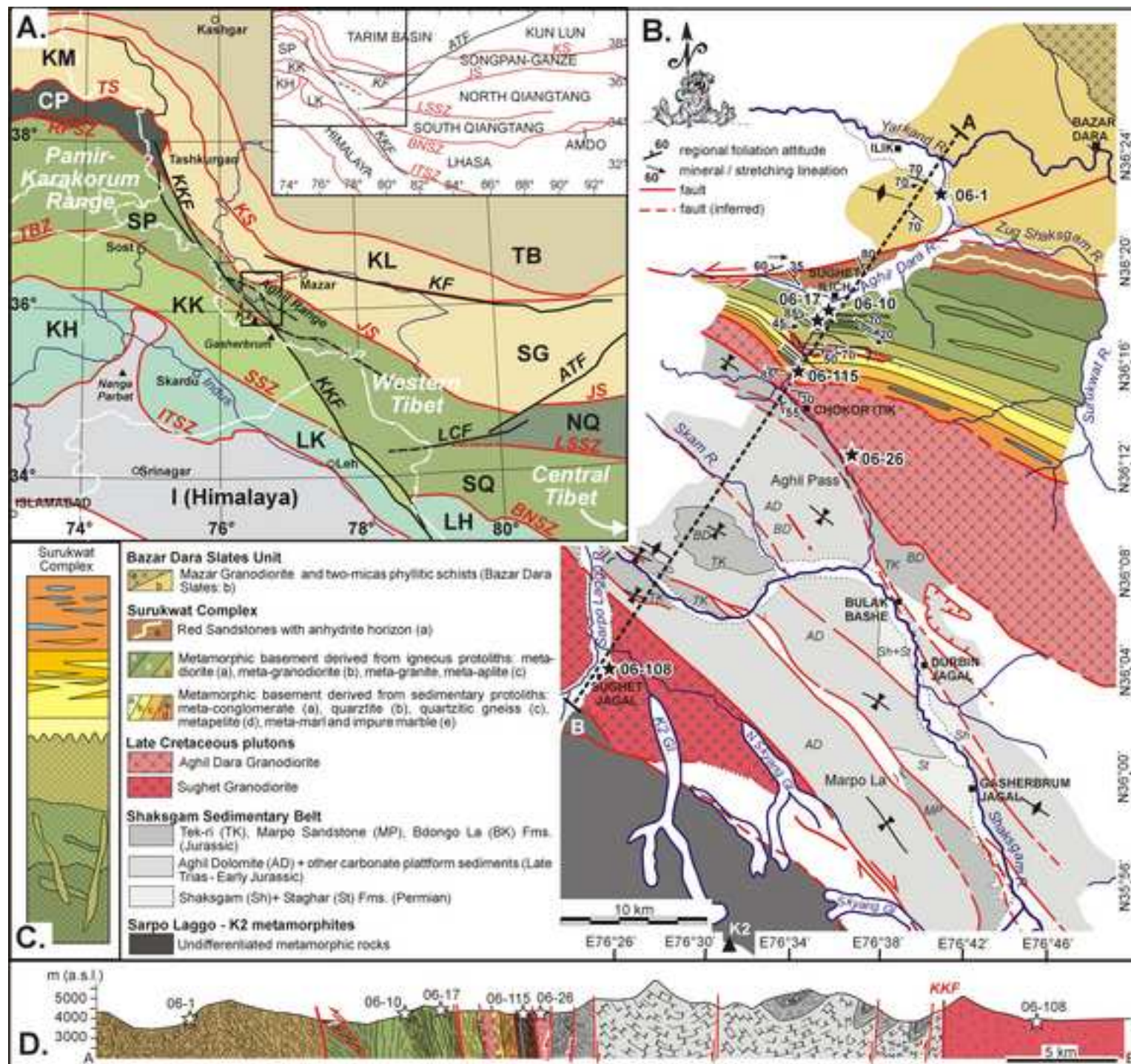




Figure 2  
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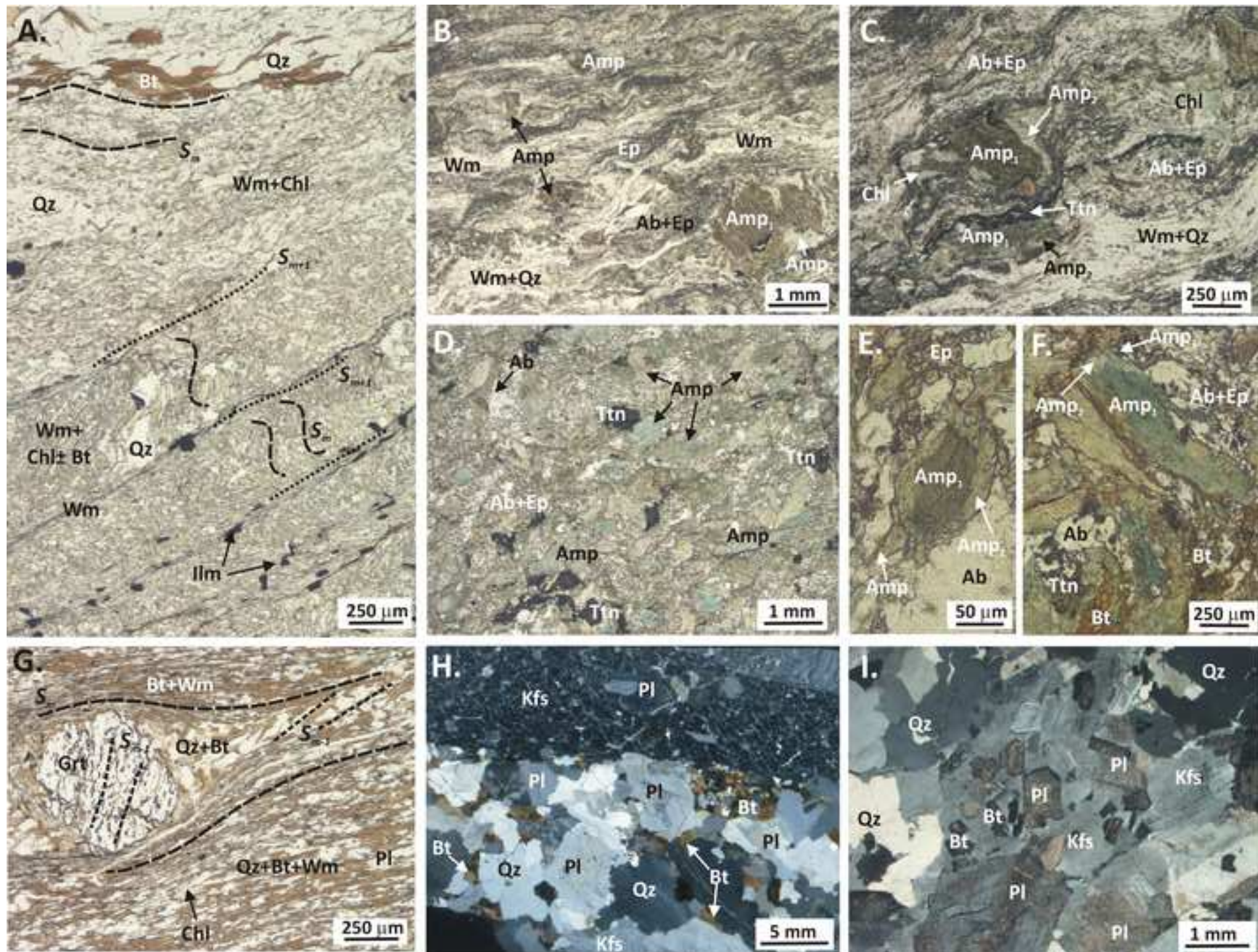




Figure 3  
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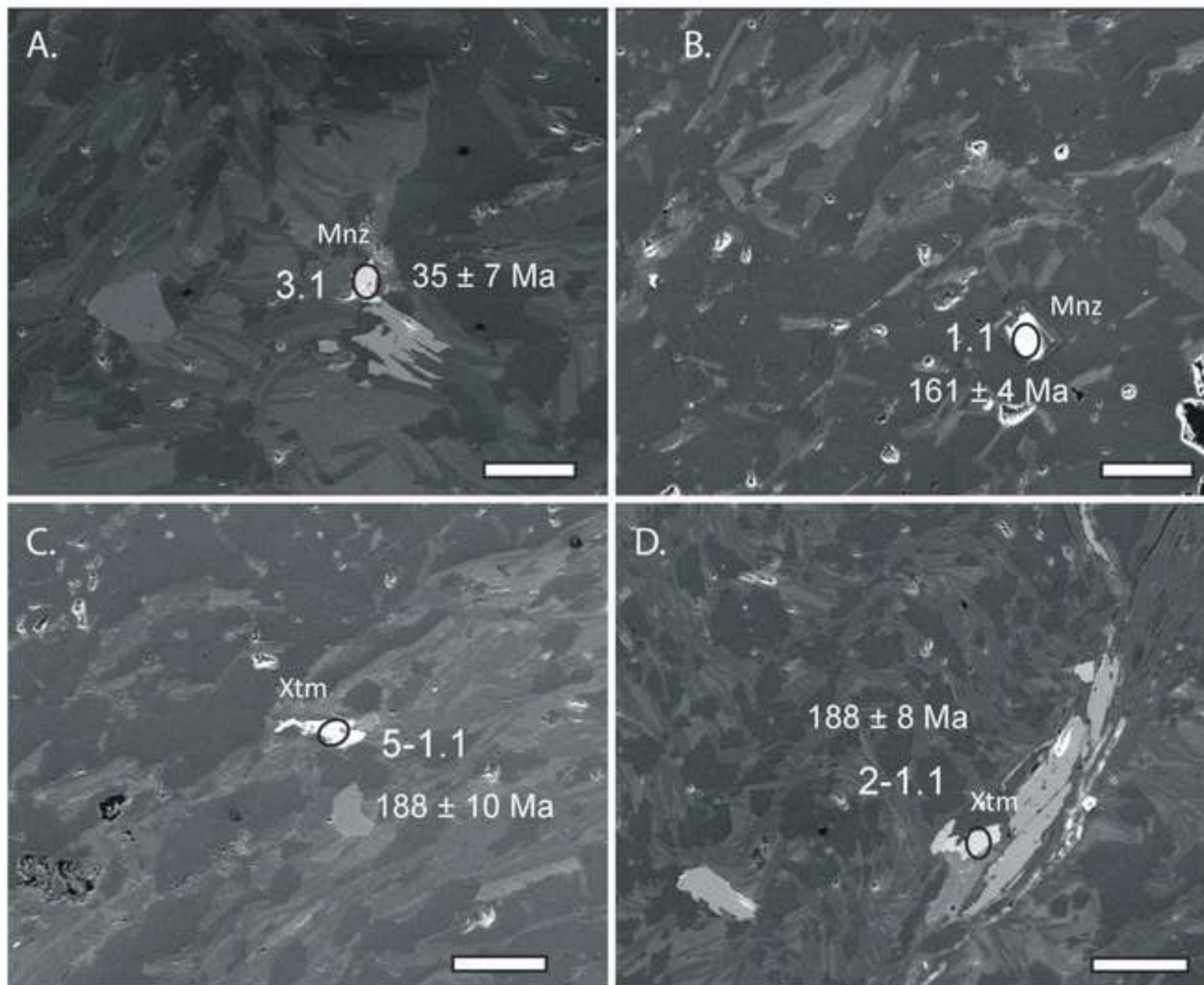


Figure 4  
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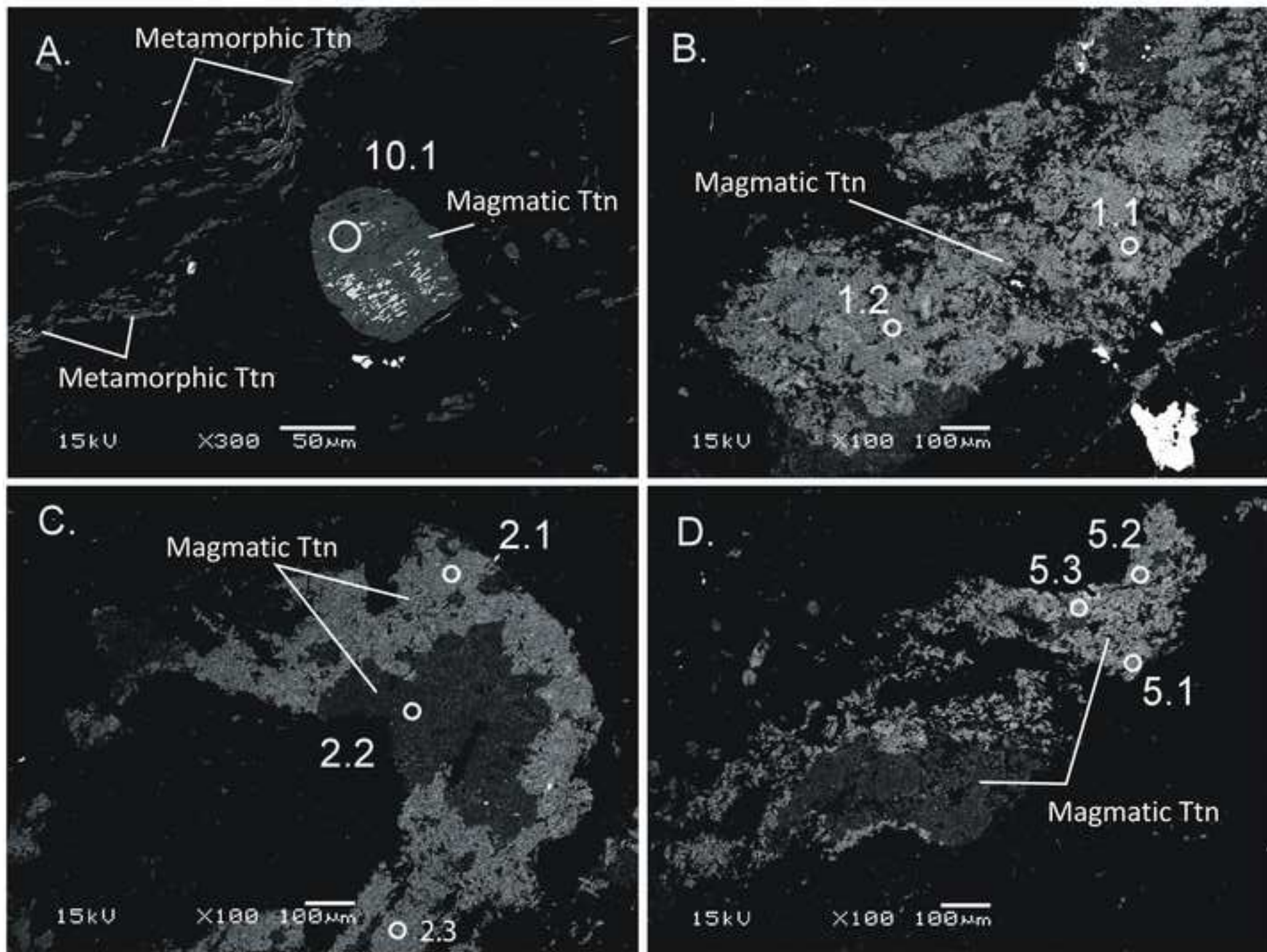




Figure 5  
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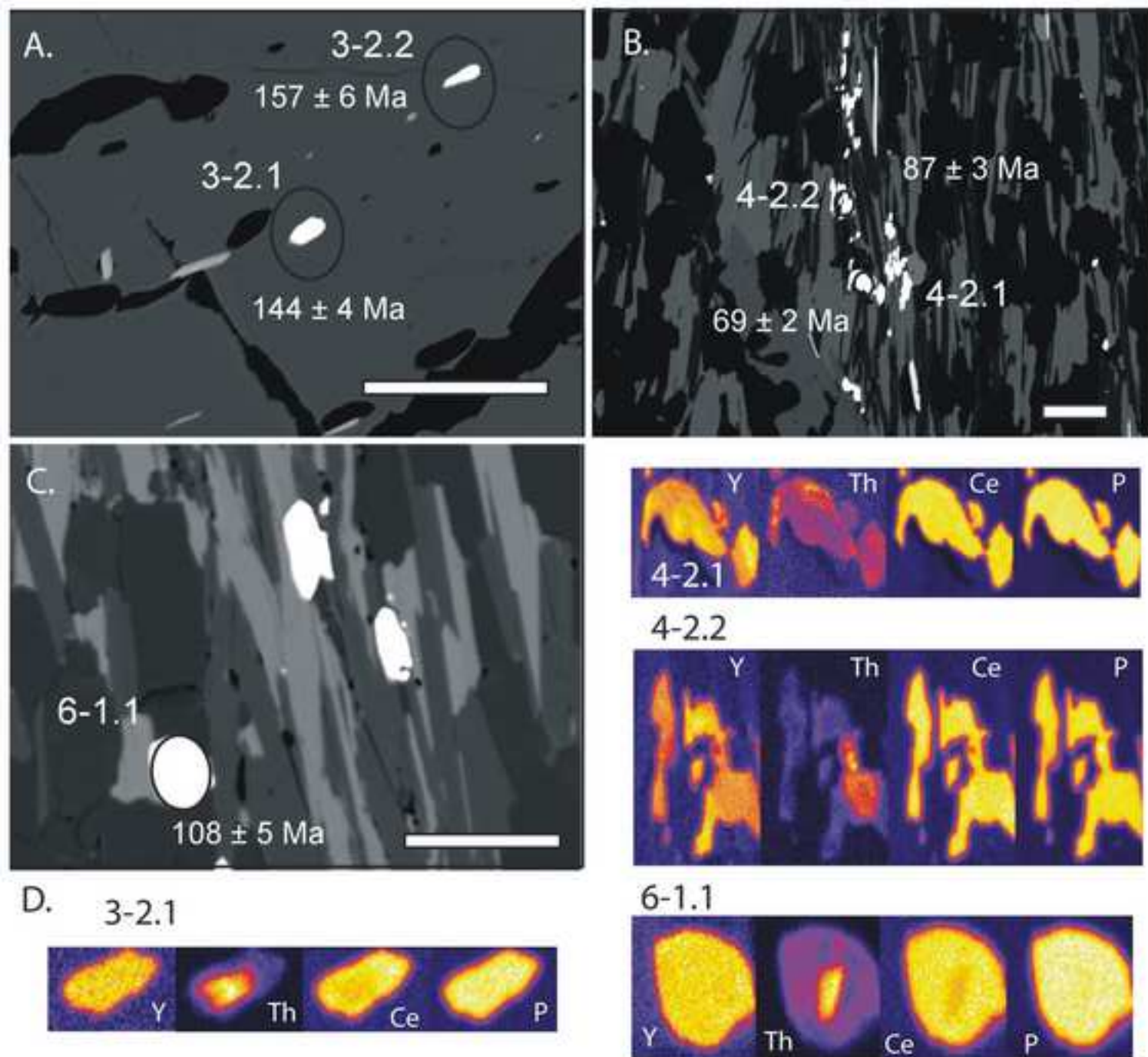


Figure 6  
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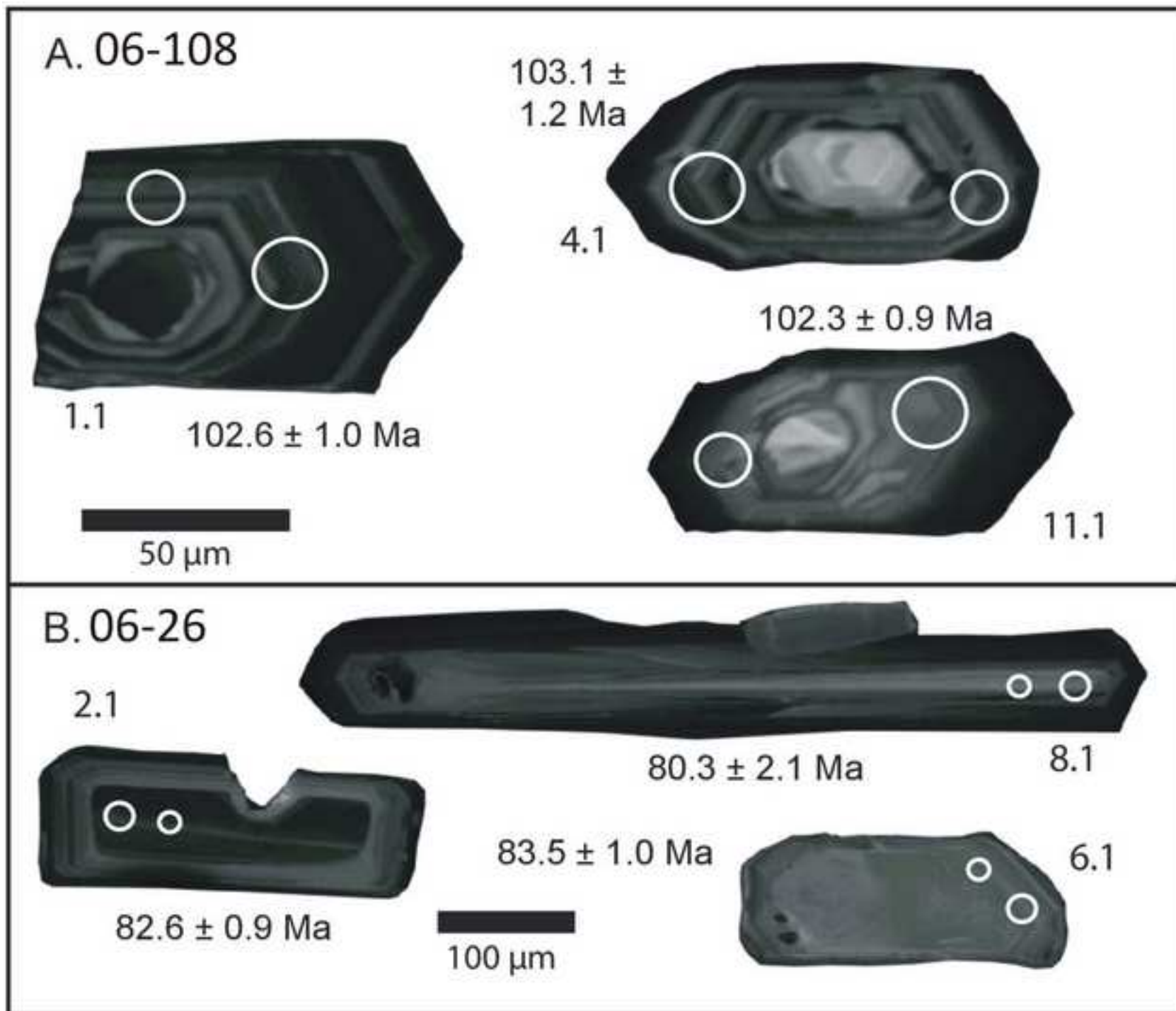


Figure 7  
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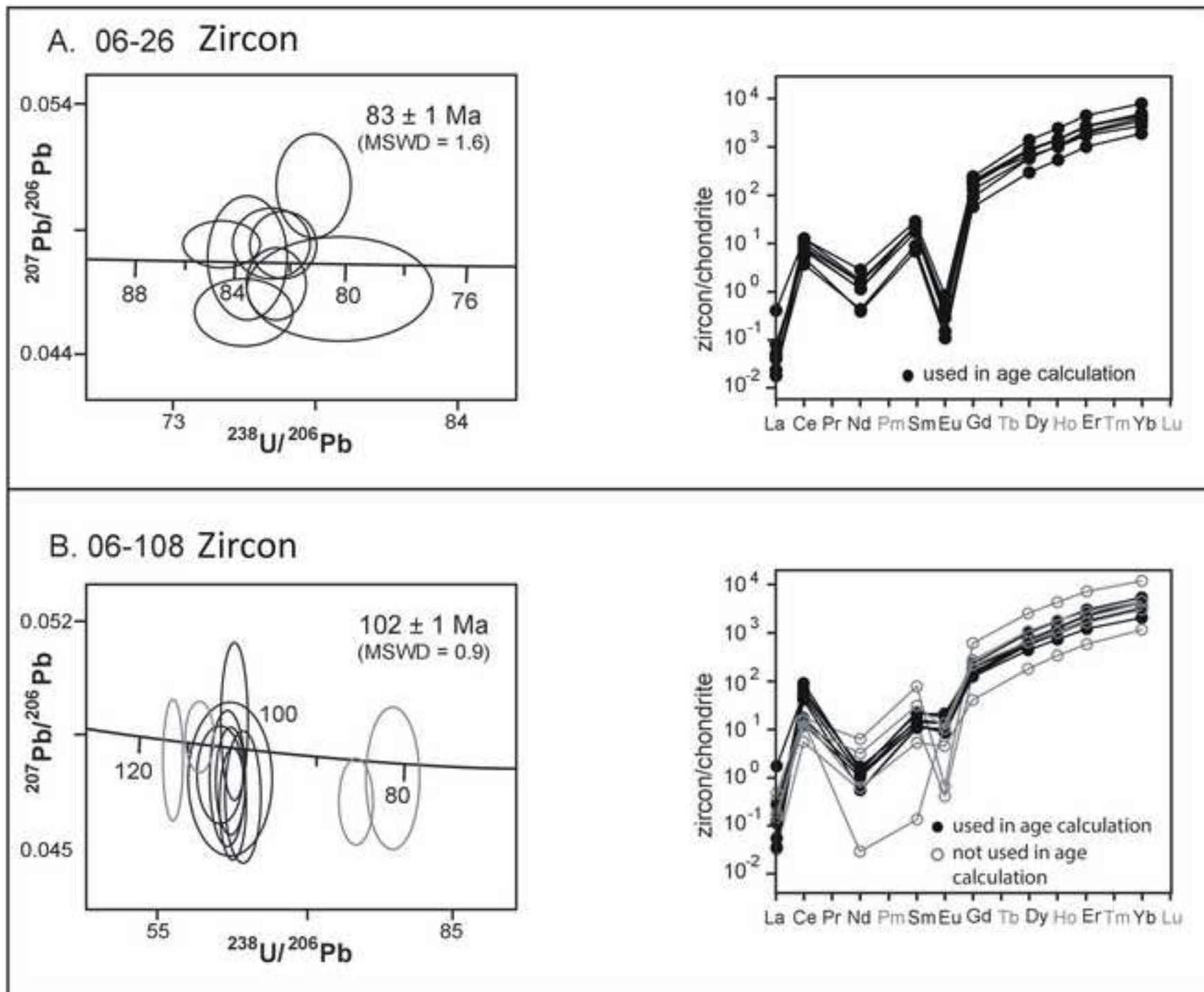


Figure 8

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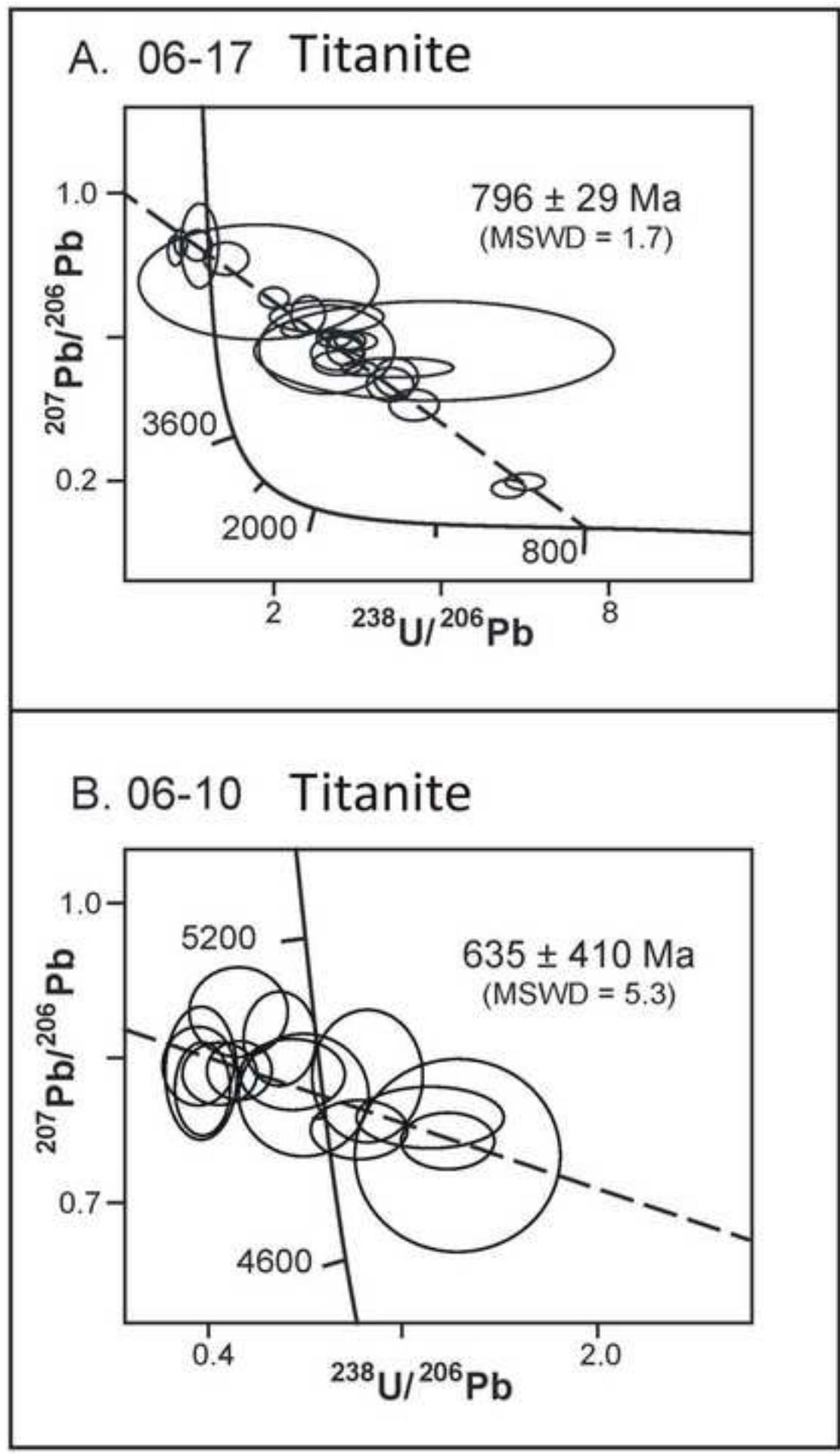




Figure 9  
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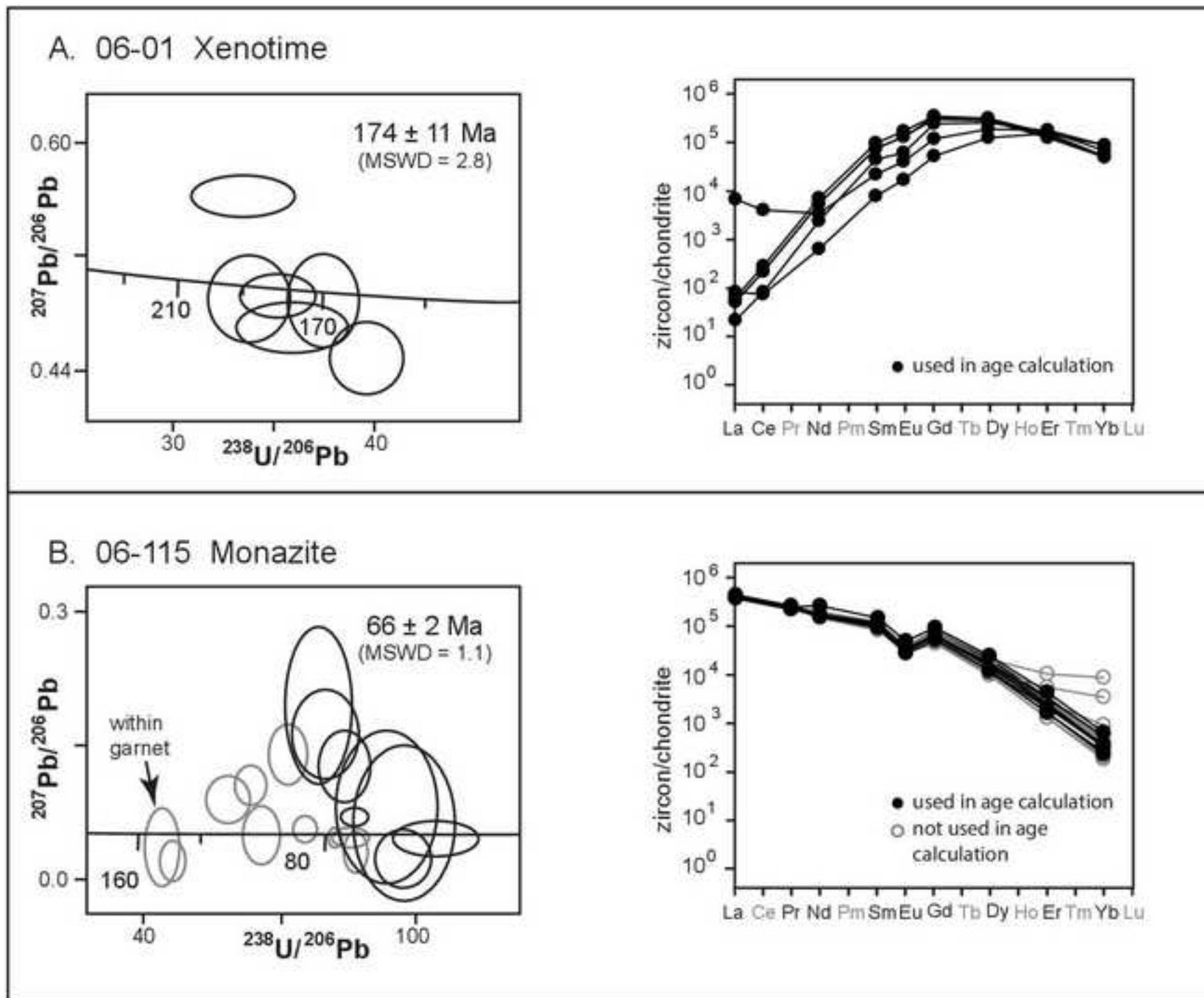




Figure 10

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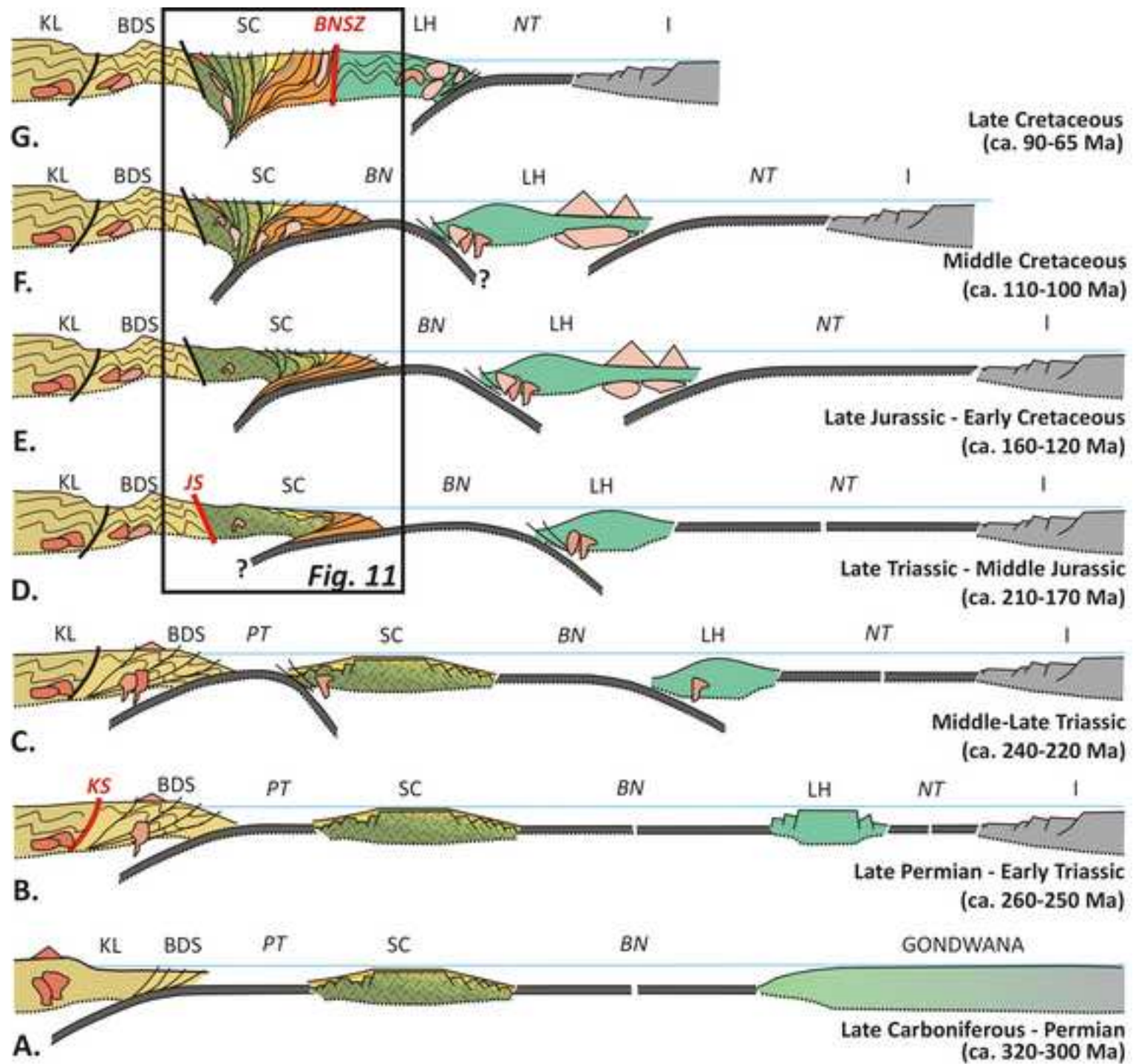


Figure 11

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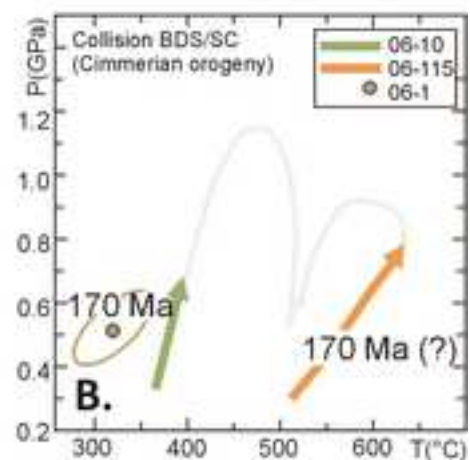
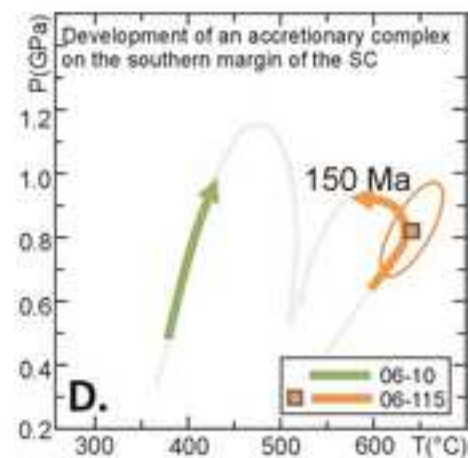
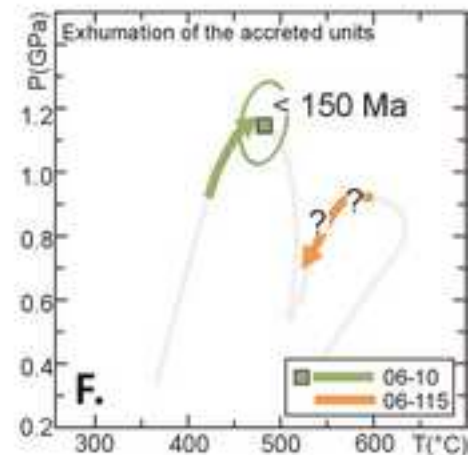
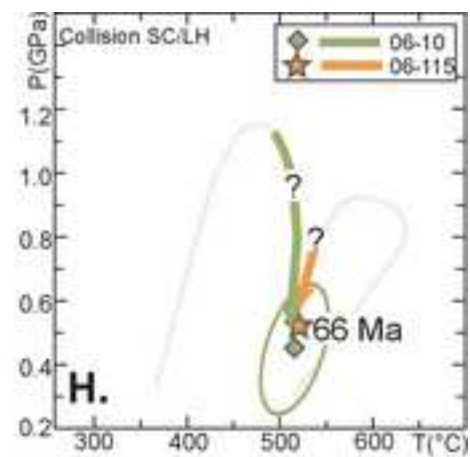
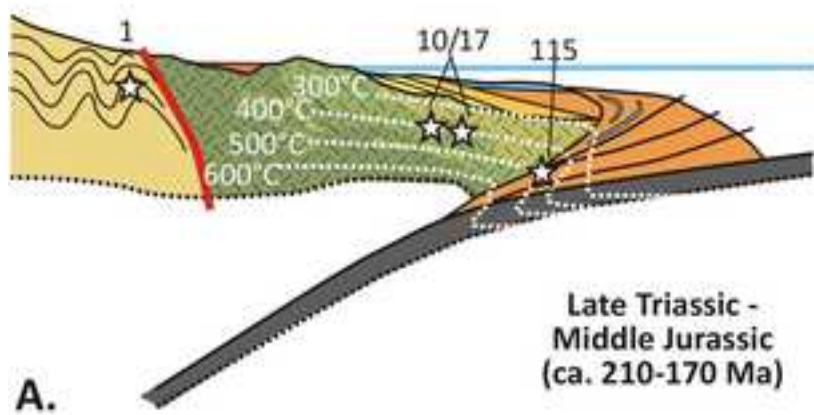
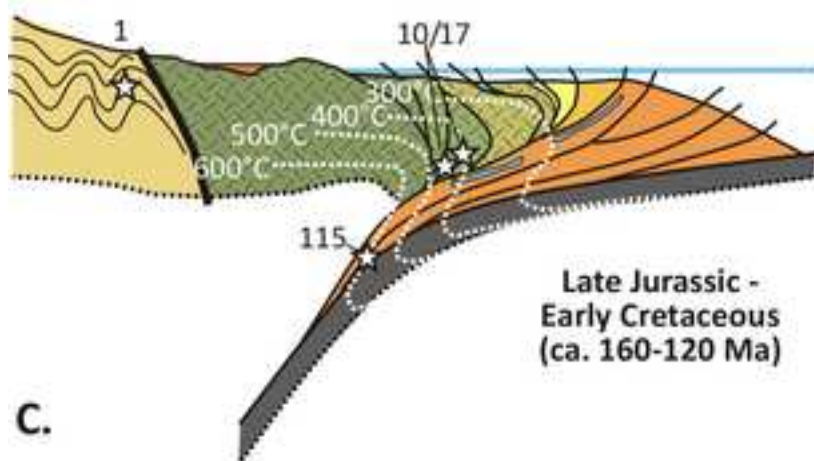
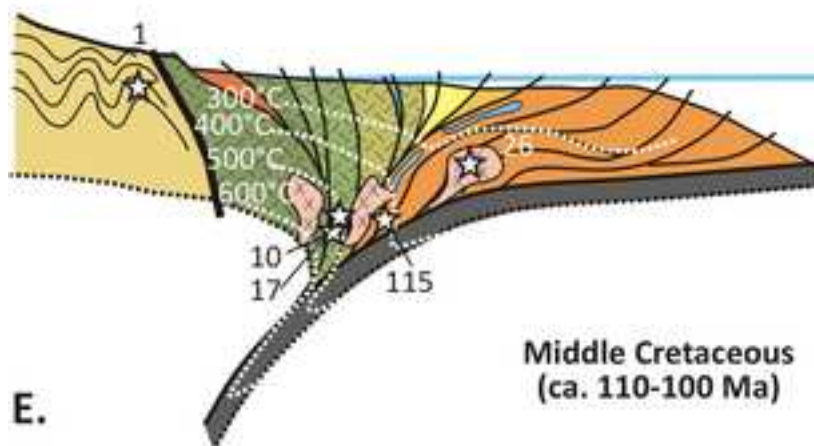
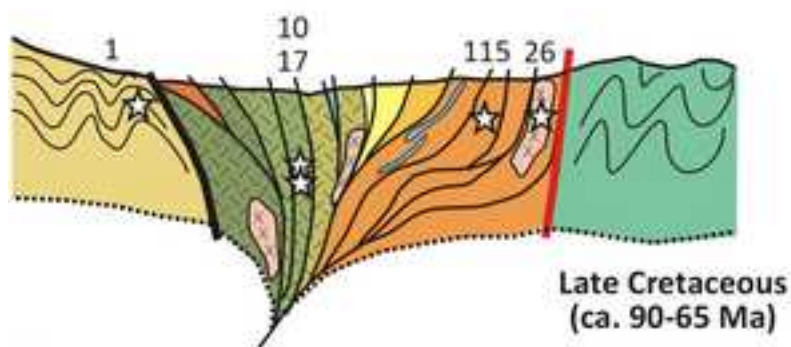




Figure 12  
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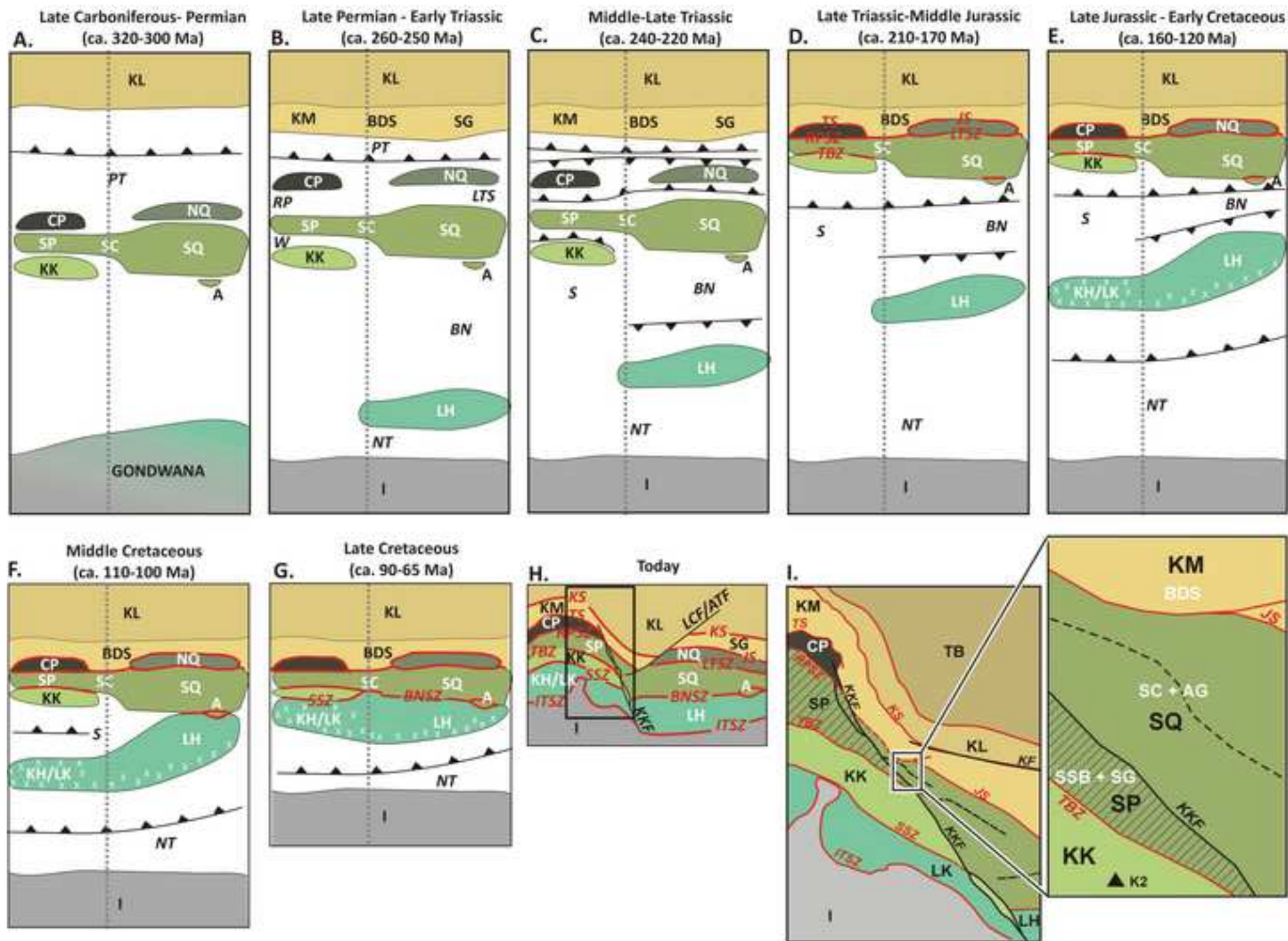


Table 1a

Table 1a.

Representative analyses of minerals in the meta-granodiorite (06-10) and metadiorite (06-17) from the Surukwat Complex.

Sample Assemblage Analyses	06-10										06-17							
	Peak-P assemblage					Peak-T assemblage					Peak-P assemblage				Peak-T assemblage			
	Act 5.4	Phe 7.13	Ep 2.11	Chl 2.12	Ab 8.4	Hbl 7.3	Mu 4.2	Ep 4.5	Chl 7.8	Ab 7.8	Act 7.2	Chl 1.10	Ep 1.9	Ab 7.6	Hbl 2.3	Bi 4.4	Ep 4.9	Pl 7.5
SiO <sub>2</sub>	53.69	49.11	37.49	27.49	68.55	49.85	47.53	38.18	28.21	67.60	52.10	29.70	38.07	68.11	42.86	36.50	37.76	67.24
TiO <sub>2</sub>	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.01	1.99	0.00	0.00
Al <sub>2</sub> O <sub>3</sub>	2.57	26.18	23.57	20.51	19.68	6.65	27.06	23.18	20.33	19.77	2.70	17.86	22.34	19.22	9.81	15.26	22.73	19.09
FeO	11.88	3.76	12.37	18.44	0.00	13.69	5.36	13.47	20.51	0.00	17.25	24.84	14.10	0.00	21.24	22.53	11.88	0.00
MnO	0.51	0.00	0.01	0.01	0.00	0.01	0.00	0.01	0.01	0.00	0.47	0.00	0.00	0.00	0.01	0.00	0.00	0.00
MgO	17.39	3.45	0.55	20.36	0.00	14.24	3.68	0.00	19.47	0.00	12.38	12.24	1.35	0.00	7.27	9.19	0.00	0.00
CaO	12.37	0.00	23.34	0.00	0.32	12.38	0.00	24.30	0.00	0.36	11.93	0.00	19.92	0.25	10.94	0.00	23.49	1.03
Na <sub>2</sub> O	0.01	0.01	0.00	0.00	12.31	1.00	0.29	0.00	0.00	12.21	0.05	0.00	0.00	11.16	1.36	0.00	0.00	11.14
K <sub>2</sub> O	0.23	10.90	0.00	0.00	0.00	0.35	10.48	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.78	9.49	0.00	0.00
Total	98.66	93.41	97.33	86.81	100.86	98.18	94.40	99.14	88.53	99.94	96.90	84.64	95.78	98.74	94.28	94.96	95.86	98.50
Si	7.577	3.366	2.968	2.809	2.978	7.193	3.239	2.978	2.851	2.965	7.707	3.198	3.043	3.005	6.746	2.836	3.030	2.986
Ti	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.001	0.116	0.000	0.000
Al	0.428	2.116	2.200	2.471	1.008	1.131	2.174	2.132	2.422	1.022	0.471	2.268	2.105	1.000	1.820	1.398	2.151	0.999
Fe <sup>+3</sup>	0.374	0.097	0.800	0.000	0.000	0.229	0.214	0.868	0.000	0.000	0.110	0.000	0.895	0.000	0.278	0.000	0.790	0.000
Fe <sup>+2</sup>	1.029	0.118	0.019	1.576	0.000	1.423	0.092	0.011	1.733	0.000	2.024	2.237	0.048	0.000	2.520	1.464	0.008	0.000
Mn	0.061	0.000	0.001	0.001	0.000	0.001	0.000	0.001	0.001	0.000	0.059	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Mg	3.658	0.352	0.065	3.100	0.000	3.062	0.374	0.000	2.932	0.000	2.729	1.964	0.161	0.000	1.705	1.064	0.000	0.000
Ca	1.871	0.000	1.980	0.000	0.015	1.914	0.000	2.031	0.000	0.017	1.891	0.000	1.706	0.012	1.845	0.000	2.020	0.049
Na	0.003	0.001	0.000	0.000	1.037	0.280	0.038	0.000	0.000	1.038	0.014	0.000	0.000	0.955	0.415	0.000	0.000	0.959
K	0.041	0.953	0.000	0.000	0.000	0.064	0.911	0.000	0.000	0.000	0.002	0.000	0.000	0.000	0.157	0.941	0.000	0.000
XMg(Fe <sub>tot</sub> )	0.72			0.66		0.65			0.63		0.56	0.47			0.38	0.42		
XAb					0.99					0.98				0.99				0.95
XPs			0.27					0.29					0.30				0.27	

Table 1b

Table 1b.

Representative analyses of minerals in the metapelites from the Bazar Dara Slates (06-1) and Surukwat Complex (06-115).

Sample	06-1				06-115			
	Assemblage				Peak-P assemblage			
Analyses	Bt 1.9	Mu 1.16	Chl 1.14	Ab 1.23	Bt 2.2	Mu 8.10	Grt 5.11	Pl 3.6
SiO <sub>2</sub>	36.03	47.99	29.02	67.15	35.37	46.37	36.10	66.01
TiO <sub>2</sub>	1.97	0.60	0.00	0.00	1.85	0.42	0.00	0.00
Al <sub>2</sub> O <sub>3</sub>	16.13	29.10	20.32	19.79	18.13	35.01	20.14	20.75
FeO	19.73	3.07	22.59	0.00	21.22	1.28	33.28	0.00
MnO	0.00	0.00	0.00	0.00	0.00	0.00	6.38	0.00
MgO	12.02	2.46	15.09	0.00	9.02	0.47	2.28	0.00
CaO	0.00	0.00	0.00	0.32	0.00	0.00	0.96	1.56
Na <sub>2</sub> O	0.00	0.45	0.00	12.24	0.00	1.23	0.00	11.66
K <sub>2</sub> O	6.91	10.48	0.00	0.00	9.25	9.06	0.00	0.00
Total	92.79	94.15	87.02	99.50	94.84	93.84	99.14	99.98
Si	2.760	3.263	3.002	2.960	2.727	3.110	2.962	2.906
Ti	0.113	0.031	0.000	0.000	0.107	0.021	0.000	0.000
Al	1.457	2.333	2.478	1.028	1.648	2.768	1.948	1.077
Fe <sup>+3</sup>	0.190	0.012	0.000	0.000	0.000	0.000	0.128	0.000
Fe <sup>+2</sup>	1.074	0.163	1.954	0.000	1.368	0.072	2.155	0.000
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.443	0.000
Mg	1.372	0.249	2.326	0.000	1.036	0.047	0.279	0.000
Ca	0.000	0.000	0.000	0.015	0.000	0.000	0.084	0.074
Na	0.000	0.059	0.000	1.046	0.000	0.160	0.000	0.995
K	0.675	0.909	0.000	0.000	0.910	0.775	0.000	0.000
XMg(Fe <sub>tot</sub> )	0.52		0.54		0.43			
XAb				0.99				0.93

**Table 2**

Average pressure-temperatures estimates for the studied samples.

Unit	Sample	Assemblage	T (°C)	P (kbar)	$\sigma_{fit}$	N° of reactions
Bazar Dara Slates	06-1	Bt-Mu-Chl-Pl-Qz-H <sub>2</sub> O	320 ± 32	5.2 ± 0.9	0.99	4
Surukwat Complex	06-10 (peak-P)	Act-Ab-Phe-Chl-Ep-Qz-Ttn-Ru-H <sub>2</sub> O	482 ± 20	11.5 ± 1.2	1.10	5
Surukwat Complex	06-10 (peak-T)	Hbl-Ab-Mu-Chl-Ep-Qz-Ttn-Ru-H <sub>2</sub> O	512 ± 30	4.5 ± 1.7	1.33	6
Surukwat Complex	06-115	Grt-Mu-Bt-Pl-Qz-Ilm-H <sub>2</sub> O	645 ± 26	8.2 ± 1.2	0.71	6

Table 3

SIMS U-Pb geochronologic zircon data and apparent ages.

Spot <sup>a</sup>	U (ppm)	Th (ppm)	Th/U	<sup>206</sup> Pb* <sup>b</sup> (ppm)	f <sup>206</sup> Pb <sub>c</sub> <sup>b</sup>	<sup>238</sup> U/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>207</sup> Pb/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>206</sup> Pb/ <sup>238</sup> U <sup>d</sup> (Ma)	1σ (%)
<i>Sample 06-108</i>											
10.1	1208	215	0.18	11	<0.01	78.911	1.4	0.04719	1.9	81.2	1.1
3.1	3679	910	0.25	37	<0.01	75.140	1.0	0.04641	1.1	85.4	0.8
12.1	1179	374	0.32	14	<0.01	63.743	1.0	0.04662	1.8	<b>101</b>	<b>1.0</b>
5.1	1001	348	0.35	12	0.11	62.778	0.8	0.04894	2.0	<b>102</b>	<b>0.8</b>
11.1	1550	398	0.26	19	<0.01	62.638	0.8	0.04641	1.5	<b>102</b>	<b>0.9</b>
1.1	1901	621	0.33	23	<0.01	62.407	1.0	0.04710	1.5	<b>103</b>	<b>1.0</b>
7.1	922	174	0.19	11	<0.01	62.284	2.8	0.04718	2.0	<b>103</b>	<b>2.9</b>
4.1	1286	468	0.36	16	<0.01	62.124	1.2	0.04717	1.8	<b>103</b>	<b>1.2</b>
6.1	2106	267	0.13	26	<0.01	61.323	1.6	0.04729	1.3	<b>104</b>	<b>1.6</b>
9.1	3976	430	0.11	50	0.03	59.251	1.0	0.04845	0.9	108	1.1
8.1	1403	306	0.22	19	<0.01	56.574	0.8	0.04772	1.6	113	0.9
2.1	223	68	0.31	52	2.46	3.134	1.7	0.12830	1.4	1747	30
<i>Sample 06-26</i>											
6.1	514	291	0.57	5	<0.01	79.892	2.7	0.04660	3.0	<b>80.3</b>	<b>2.1</b>
4.1	782	109	0.14	7	0.39	78.969	1.1	0.05072	2.7	<b>80.8</b>	<b>0.9</b>
3.1	1355	281	0.21	13	0.09	77.912	1.0	0.04838	1.8	<b>82.1</b>	<b>0.8</b>
5.1	1182	216	0.18	11	<0.01	77.658	0.9	0.04682	2.0	<b>82.6</b>	<b>0.7</b>
2.1	1323	324	0.24	13	0.09	77.470	1.1	0.04844	1.9	<b>82.6</b>	<b>0.9</b>
8.1	1320	296	0.22	13	0.02	76.660	1.2	0.04784	3.5	<b>83.5</b>	<b>1.0</b>
7.1	1138	124	0.11	11	<0.01	76.540	1.5	0.04571	2.0	<b>83.9</b>	<b>1.2</b>
1.1	2451	363	0.15	24	0.09	75.744	1.2	0.04842	1.3	<b>84.5</b>	<b>1.0</b>

<sup>a</sup> Label format is grain number.spot number<sup>b</sup> Pb\* denotes radiogenic Pb; Pb<sub>c</sub> denotes common Pb;  $f^{206}\text{Pb}_c = 100 * (\frac{^{206}\text{Pb}_c}{^{206}\text{Pb}_{\text{total}}})$ <sup>c</sup> Calibration concentrations and isotopic compositions were based on replicate analyses of R33 (419 Ma, Black et al., 2004) and Madagascar Green (MADDER; 3435 ppm U, Barth & Wooden, 2010). Reported ratios are not corrected for common Pb. Errors are reported as percent at the 1σ level.<sup>d</sup> Ages were calculated from <sup>206</sup>Pb/<sup>238</sup>U ratios corrected for common Pb using the <sup>207</sup>Pb method and <sup>207</sup>Pb/<sup>206</sup>Pb ratios corrected for common Pb using the <sup>204</sup>Pb method (see Williams, 1998). Initial common Pb isotopic composition approximated from Stacey & Kramers (1975). Uncertainties in millions of years reported as 1σ. Apparent ages in bold are used in age calculations discussed in the text.

Table 4

SIMS U-Pb geochronologic titanite data and apparent ages.

Spot <sup>a</sup>	<sup>206</sup> Pb/ <sup>204</sup> Pb	1σ	U	Th	Th/U	<sup>206</sup> Pb* <sup>b</sup>	f <sup>206</sup> Pb <sub>c</sub> <sup>b</sup>	<sup>238</sup> U/ <sup>206</sup> Pb <sup>c</sup>	1σ	<sup>207</sup> Pb/ <sup>206</sup> Pb <sup>c</sup>	1σ	<sup>206</sup> Pb/ <sup>238</sup> U <sup>d</sup>	1σ	<sup>207</sup> Pb/ <sup>206</sup> Pb <sup>d</sup>	1σ
		(%)	(ppm)	(ppm)		(ppm)			(%)		(%)	(Ma)	(%)	(Ma)	(%)
<i>Sample 06-10</i>															
3.2	20.2	13	0.1	0.1	0.65	0.00	98.47	0.5	15.9	0.90	2.1	186	266	5,132	41
2.2	37.9	12	0.6	0.1	0.17	0.04	95.22	0.7	8.8	0.87	2.3	430	209	3,994	2,312
1.2	35.8	14	0.4	0.1	0.26	0.03	90.68	1.1	8.9	0.83	3.4	546	193	5,951	3,220
11.1	44.1	16	0.2	0.1	0.26	0.02	90.78	0.7	11.8	0.83	1.8	753	164	4,367	599
3.1	22.5	10	0.7	0.1	0.11	0.07	85.58	1.3	9.5	0.79	1.6	671	93	5,412	82
2.1	46.6	13	0.4	0.0	0.08	0.05	91.20	0.5	10.4	0.84	1.5	998	191	4,558	274
8.1	20.1	10	0.4	0.0	0.13	0.04	88.22	0.8	13.8	0.81	3.2	895	256	5,300	65
6.2	49.0	13	1.8	0.1	0.08	0.20	82.71	1.4	5.7	0.76	1.6	758	73	3,939	606
1.1	46.6	20	0.3	0.1	0.25	0.03	80.88	1.4	12.2	0.75	5.4	811	220	3,358	2,211
5.1	49.4	18	0.1	0.0	0.14	0.03	91.62	0.4	17.0	0.84	2.0	1,365	389	4,683	245
6.1	18.0	9	0.2	0.0	0.23	0.03	90.73	0.4	14.2	0.83	1.6	1,221	262	5,227	29
6.3	52.5	13	0.5	0.1	0.30	0.06	83.95	1.0	8.0	0.77	1.6	942	111	4,275	306
5.2	44.2	15	0.2	0.0	0.09	0.04	90.66	0.4	16.6	0.83	3.4	1,459	571	4,393	607
10.1	40.5	21	0.3	0.1	0.23	0.08	88.48	0.4	12.5	0.82	2.4	1,733	422	3,248	5,010
<i>Sample 06-17</i>															
5.2	43.6	16	5.0	5.0	1.04	0.36	59.07	4.9	26.7	0.56	10.1	517	158	4,929	596
3.2	30.6	18	0.2	0.0	0.19	0.01	93.54	0.7	18.9	0.85	5.6	554	482	5,628	762
5.1	41.6	19	0.2	0.2	0.95	0.02	89.25	1.2	13.5	0.82	2.3	554	132	3,747	2,271
2.3	35.4	15	0.2	0.1	0.33	0.01	93.53	0.7	15.1	0.85	2.0	582	199	6,030	5,146
5.3	47.4	15	2.2	3.0	1.41	0.19	70.25	3.0	14.0	0.66	2.8	613	94		
6.1	67.1	18	6.3	2.8	0.47	0.56	81.49	1.8	49.2	0.75	8.7	642	402	4,469	371
7.1	68.2	16	2.1	2.4	1.16	0.20	62.03	3.5	4.3	0.59	1.8	663	35	3,208	702
13.2	58.7	15	2.3	4.5	2.00	0.22	70.86	2.7	4.7	0.66	3.2	672	64	3,496	700
3.1	54.9	17	1.9	1.7	0.91	0.18	53.34	4.2	9.6	0.51	2.2	672	65	2,525	1,666
12.1	68.3	17	2.3	2.4	1.08	0.22	62.20	3.3	4.2	0.59	1.7	692	35	3,241	704
2.1	38.0	13	1.3	0.9	0.75	0.13	76.33	2.0	5.3	0.71	1.7	705	54	5,725	1,070
12.2	64.4	15	2.4	3.3	1.43	0.24	50.07	4.2	3.8	0.49	4.5	721	45		
12.1	62.0	15	2.2	2.3	1.10	0.22	62.84	3.1	4.0	0.60	1.6	727	36	2,789	1,168
12.3	64.1	17	2.5	3.6	1.52	0.26	59.28	3.3	4.0	0.57	3.2	748	48	2,445	1,651
2.2	64.1	17	2.1	1.7	0.85	0.23	47.62	4.1	4.1	0.47	4.1	773	45	433	6,383
8.3	258.1	22	10.6	12.8	1.24	1.19	15.47	6.5	2.2	0.20	4.7	786	20	1,319	681
1.2	77.3	19	6.2	3.5	0.59	0.69	40.86	4.5	4.0	0.41	4.3	790	41		
9.1	76.3	19	1.8	2.1	1.22	0.20	57.97	3.2	6.0	0.55	3.5	799	63	3,243	671
7.2	69.2	14	2.8	2.8	1.02	0.32	52.89	3.6	3.6	0.51	1.6	801	32	1,839	1,775
1.1	37.9	14	0.4	0.0	0.10	0.05	94.60	0.4	11.9	0.86	1.7	825	275	3,778	3,485
8.2	46.0	15	1.1	1.0	1.01	0.12	59.09	3.0	16.5	0.56	8.9	828	176	4,218	629
4.2	268.4	22	9.8	6.4	0.68	1.17	13.00	6.2	1.9	0.18	5.6	844	19	867	910
13.1	65.0	14	2.6	2.3	0.92	0.31	65.83	2.4	4.2	0.62	1.5	848	44	3,429	534
8.1	43.0	19	2.1	2.0	1.02	0.25	54.52	3.2	5.9	0.53	2.6	857	57	5,333	693
4.1	38.1	16	0.1	0.0	0.29	0.03	91.47	0.3	18.6	0.84	2.0	1,646	495	2,164	16,887

<sup>a</sup> Label format is grain number.spot number<sup>b</sup> Pb\* denotes radiogenic Pb; Pb<sub>c</sub> denotes common Pb; f<sup>206</sup>Pb<sub>c</sub> = 100\*(<sup>206</sup>Pb/<sup>206</sup>Pb<sub>total</sub>)<sup>c</sup> Calibration concentrations and isotopic compositions were based on replicate analyses of MMs titanite (524 Ma; Schoene & Bowring, 2006) and BLR titanite (Aleinikoff et al., 2007). Reported ratios are not corrected for common Pb. Errors are reported as percent at the 1σ level.<sup>d</sup> Common Pb isotopic composition determined by 3D linear regression on a Tera-Wasserburg plot which gave a data-defined <sup>207</sup>Pb/<sup>206</sup>Pb upper intercept of 0.9096. Ages were calculated from <sup>206</sup>Pb/<sup>238</sup>U ratios corrected for common Pb using the <sup>207</sup>Pb method and <sup>207</sup>Pb/<sup>206</sup>Pb ratios corrected for common Pb using the <sup>204</sup>Pb method (see Williams, 1998). Uncertainties in millions of years reported as 1σ.



Table 5

SIMS U-Pb geochronologic monazite data and apparent ages.

Spot <sup>a</sup>	<sup>206</sup> Pb/ <sup>204</sup> Pb	1σ (%)	U (ppm)	Th (ppm)	Th/U	<sup>206</sup> Pb* <sup>b</sup> (ppm)	f <sup>206</sup> Pb <sub>c</sub> <sup>b</sup>	<sup>238</sup> U/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>207</sup> Pb/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>206</sup> Pb/ <sup>238</sup> U <sup>d</sup> (Ma)	1σ (%)
<i>Sample 06-115</i>													
1-1.1	35	12	947	5290	5.8	6.6	50.01	62	2.6	0.443	4.7	52	3
5-1.1	262	19	1391	5388	4.0	11.6	6.77	96	4.5	0.101	4.9	<b>62</b>	<b>3</b>
2-2.1	64	12	1279	5535	4.5	10.5	42.11	61	4.5	0.381	9.3	<b>61</b>	<b>5</b>
5-2.1	139	27	1895	31614	17.2	16.3	9.05	91	3.1	0.119	8.8	<b>64</b>	<b>2</b>
4-3.1	79	13	1754	4340	2.6	15.0	34.59	66	3.8	0.321	4.5	<b>64</b>	<b>3</b>
8-3.1	85	22	2238	10160	4.7	19.4	24.93	75	4.9	0.245	10.1	<b>65</b>	<b>4</b>
8-4.1	72	8	1578	3031	2.0	14.0	23.53	74	3.1	0.234	1.3	<b>66</b>	<b>2</b>
2-1.1	52	8	1272	2455	2.0	11.4	37.21	60	4.5	0.342	8.5	<b>67</b>	<b>5</b>
4-1.1	449	19	1211	1895	1.6	11.2	6.70	86	2.6	0.100	2.2	<b>69</b>	<b>2</b>
10-2.1	72	10	1289	2737	2.2	12.0	33.30	62	3.0	0.311	2.6	<b>69</b>	<b>2</b>
12-2.1	82	8	1110	3617	3.4	10.9	18.07	72	1.7	0.191	5.0	73	2
9-2.1	149	12	1277	2956	2.4	12.6	10.47	78	1.7	0.131	3.2	74	1
9-1.1	320	13	1343	4684	3.6	13.4	5.54	81	2.6	0.091	3.4	74	2
8-1.1	490	18	1877	5349	2.9	19.1	3.46	81	1.9	0.075	2.0	76	1
8-2.1	93	11	1063	2861	2.8	10.8	29.29	60	2.9	0.280	3.5	75	3
12-1.1	175	10	1131	2587	2.4	12.1	11.56	71	1.7	0.139	1.5	80	1
4-2.1	143	12	1904	12440	6.8	22.2	19.09	60	3.0	0.199	3.1	87	3
10-1.1	93	10	1702	24941	15.1	21.9	20.05	53	2.1	0.207	3.3	96	2
6-1.1	117	12	1537	45229	30.4	22.3	20.15	47	4.4	0.208	1.8	108	5
3-1.1	205	18	810	39544	50.5	15.7	5.48	42	3.1	0.093	3.1	144	4
3-2.1	82	17	907	26904	30.6	19.2	14.41	35	3.6	0.164	3.5	157	6
<i>Sample 06-1</i>													
3.1	37	25	4419	28129	6.6	20.7	50.96	90	49.1	0.450	2.3	35.1	17
2.1	455	30	199	3481	18.0	4.1	2.63	41	1.8	0.070	10.7	153	3
1.1	550	32	265	4766	18.6	5.7	1.81	39	2.8	0.064	4.0	161	4

<sup>a</sup> Label format is grain number.spot number<sup>b</sup> Pb\* denotes radiogenic Pb; Pb<sub>c</sub> denotes common Pb; f<sup>206</sup>Pb<sub>c</sub> = 100\*(<sup>206</sup>Pb<sub>c</sub>/<sup>206</sup>Pb<sub>total</sub>)<sup>c</sup> Calibration concentrations and isotopic compositions were based on replicate analyses of 44069 monazite (425 Ma; Aleinikoff et al., 2006, 2012). Reported ratios are not corrected for common Pb. Errors are reported as percent at the 1σ level.<sup>d</sup> Ages were calculated from <sup>206</sup>Pb/<sup>238</sup>U ratios corrected for common Pb using the <sup>207</sup>Pb method and <sup>207</sup>Pb/<sup>206</sup>Pb ratios corrected for common Pb using the <sup>204</sup>Pb method (see Williams, 1998). Initial common Pb isotopic composition approximated from Stacey & Kramers (1975). Uncertainties in millions of years reported as 1σ. Apparent ages in bold are used in age calculations discussed in the text.

**Table 6**

SIMS U-Pb geochronologic xenotime data and apparent ages.

Spot <sup>a</sup>	U (ppm)	Th (ppm)	Th/U	<sup>206</sup> Pb* <sup>b</sup> (ppm)	$f^{206}\text{Pb}_c^b$	<sup>238</sup> U/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>207</sup> Pb/ <sup>206</sup> Pb <sup>c</sup>	1σ (%)	<sup>206</sup> Pb/ <sup>238</sup> U <sup>d</sup> (Ma)	1σ (%)
<i>Sample 06-1</i>											
3-1.3	6640	6612	1.03	-0.430	145	39.5	3.0	0.0459	2.9	162	5
4-1.1	4053	1707	0.44	0.097	93	37.4	3.1	0.0503	3.4	170	5
3-1.1	4687	3818	0.84	-0.335	112	35.9	5.2	0.0470	2.5	177	9
3-1.2	5450	4493	0.85	-0.058	133	35.2	3.5	0.0493	2.1	181	6
2-1.1	10579	754	0.07	0.106	269	33.7	4.1	0.0507	2.9	188	8
5-1.1	19116	16396	0.89	0.877	487	33.4	5.2	0.0569	1.4	188	10
<i>Sample 06-115</i>											
10-1.1	8393	4994	0.61	16.617	156	38.6	3.9	0.1811	3.3	139	8

<sup>a</sup> Label format is grain number.spot number<sup>b</sup> Pb\* denotes radiogenic Pb; Pb<sub>c</sub> denotes common Pb;  $f^{206}\text{Pb}_c = 100 * ({}^{206}\text{Pb}_c / {}^{206}\text{Pb}_{\text{total}})$ <sup>c</sup> Calibration concentrations and isotopic compositions were based on replicate analyses of MG-1 xenotime (490 Ma; Fletcher et al., 2010; Aleinikoff et al., 2012). Reported ratios are not corrected for common Pb. Errors are reported as percent at the 1σ level.<sup>d</sup> Ages were calculated from <sup>206</sup>Pb/<sup>238</sup>U ratios corrected for common Pb using the <sup>207</sup>Pb method and <sup>207</sup>Pb/<sup>206</sup>Pb ratios corrected for common Pb using the <sup>204</sup>Pb method (see Williams, 1998). Initial common Pb isotopic composition approximated from Stacey & Kramers (1975). Uncertainties in millions of years reported as 1σ.

**Supplementary Material Figure SM1**

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