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Médard Thiry, Christine Franke, Kouakou Yao, Adam Szuszkiewicz, Carles Fàbrega, et al.. Albitization and oxidation of Variscan granitoid rocks related to the post-Variscan paleosurface in the Sudetes (Bohemian Massif, SW Poland). *International Journal of Earth Sciences*, 2022, <10.1007/s00531-022-02274-2>. <hal-03902388>

HAL Id: hal-03902388

<https://minesparis-psl.hal.science/hal-03902388v1>

Submitted on 3 Sep 2024

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HAL Authorization

International Journal of Earth Sciences

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--Manuscript Draft--

Manuscript Number:	IJES-D-22-00161R1
Full Title:	Albitization and oxidation of Variscan granitoid rocks related the to Post-Variscan paleosurface in the Sudetes (Bohemian Massif, SW Poland)
Article Type:	Original Paper
Keywords:	albitization; oxidation; granitoid; dating; paleosurface; Post-Variscan; Poland
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Order of Authors Secondary Information:	
Funding Information:	
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Post-Variscan ages. In this context, the alterations are related to regionally widespread Post-Variscan paleosurface. Weathering profiles formed in phreatic groundwater environments downgradient of highlands that provided a hydrological head. Sodium supply was likely to origin from the gigantic salt playas that characterized Permian and Triassic lowlands. Weathering was interrupted by an advance of the Mesozoic transgression. The recognition of these weathering profiles over extensive parts of the Variscan Belt provides the basis for reconstructing the Post-Variscan paleosurface and constraining Triassic and Post-Triassic geodynamics, including erosion rates and geomorphological evolution of the Paleozoic massifs in Europe.

1 **Albitization and oxidation of Variscan granitoid rocks related to the **Post-Variscan** paleosurface in the**
2 **Sudetes (Bohemian Massif, SW Poland)**

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20
21 **Abstract:**

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25 fracture walls. These alterations are associated with chloritization of primary ferromagnesian minerals and the
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27 these parageneses and the observed zonation to a unique alteration event consisting of an interplay of chemical
28 reactions of variable spatial extent. The complete albitization of the feldspars (plagioclase and K-feldspar)
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36 weathering profiles over extensive parts of the Variscan Belt provides the basis for reconstructing the **Post-**
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38 geomorphological evolution of the Paleozoic massifs in Europe.

39
40 **Keywords:**

41 albitization, oxidation, granitoid, dating, paleosurface, Triassic, Poland.

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Running title:

Albitization related to the Post-Variscan paleosurface

Statements and declarations:

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements:

This work is the result of a quest that was started more than 40 years ago at the Ecole des Mines de Paris. The French team likes to acknowledge the two participating Polish Institutions for the organization of the field work campaigns and for two scientific stays of K.Y. in the labs with access to CL. We acknowledge the IPGP Paleomagnetism group for access and help with 2G cryogenic magnetometer paleomagnetic and rock magnetic measurements as well as the CLIMAG group at the LSCE for access to the rock magnetic measurements performed in their laboratories. The authors thank Jean-Bernard Edel from the University of Strasbourg and an anonymous reviewer for their thoughtful comments, which significantly strengthened the earlier version of the article. Authors are also indebted to Tony Milnes from University of Adelaide for his discussions and language edition.

62

63 1 INTRODUCTION

64 In crystalline rocks, albitization is commonly considered to be a deep and high-temperature
65 metasomatic alteration linked to granite cooling during exhumation (e.g. Putnis et al. 2007; Engvik et al. 2008;
66 Sandström et al. 2010; Morad et al. 2010). Besides, in basaltic rocks, albitization with typical albite, chlorite, and
67 epidote assemblage is considered to be related to pervasive alteration by deep circulation of heated seawater
68 (spilitization) under relative lower temperature (Kelley et al. 1992; Alt et al. 2010; Richter and Diamond 2022).
69 Studies of albitization in granitic and other basement rocks have been mostly undertaken by petrologists who
70 usually consider its petrographical aspects and recognise the late-stage nature of the parageneses. However, the
71 geometric aspects of the albitization phenomenon have never been thoroughly investigated. In this paper we
72 describe the albitization as a result of superficial weathering, spatially and temporally associated with the Post-
73 Variscan paleosurface, by analogy with situations described locally in the French Massif Central (Yerle and
74 Thiry 1979; Schmitt 1986, 1992; Schmitt and Clement 1989; Parcerisa et al. 2009) and in NE Spain (Fàbrega et
75 al. 2019).

76 We focus our study on the Polish Sudetes where paleomagnetic dating of Post-Variscan
77 remagnetization overprints of the Paleozoic crystalline basement has already been well-reported in the literature
78 (e.g. Edel et al. 1997). Our aim is to sort out whether the albitization observed in the Sudetic granitoids could
79 have been of superficial (a weathering phenomenon) rather than of deep-seated hydrothermal origin. For this
80 purpose, we describe detailed mineralogical and petrographical features of the albitized facies rocks. Special
81 attention is paid to the petrographic evolution of the granitoids in relation to the reddened facies adjacent to
82 fractures. The relationship between primary and secondary mineral phases is emphasized with the focus to
83 establish a specific mineral paragenesis and geochemical conceptual model for the albitization, and to understand
84 its association with the contemporaneous formation of hematite. Subsequently, the results of paleomagnetic
85 dating of iron oxides and U-Th-Pb_{total} dating of monazites in the reddened facies are presented in order to explore
86 their respective links with the Post-Variscan paleosurface.

87 2 GEOLOGICAL SETTING

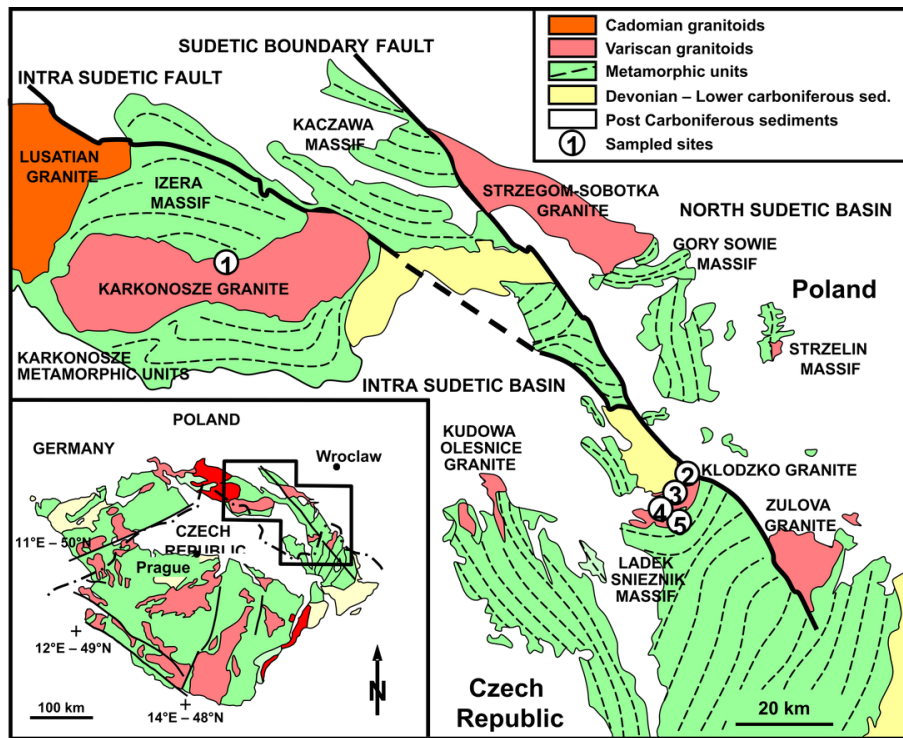
88 2.1 Regional frame

89 The mountain range of the Sudetes is the topographic expression of a regional-size fault block, about
90 300 km long and 50 to 80 km wide, at the NE margin of the Bohemian Massif (Fig. 1). The Sudetic block was
91 uplifted, together with the entire Bohemian Massif, in Late Cenozoic times along major NW-SE and NE-SW
92 trending faults at the foreland of the Eastern Alps and Carpathians (e.g. Reicherter et al. 2008), while to the NE it
93 is accompanied by the geologically similar Fore-Sudetic block that is currently not elevated.

94 The Sudetes occur on the northeastern periphery the internal zone of the Variscan belt of Europe, whose
95 mostly crystalline, Late Neoproterozoic to Carboniferous basement rock complexes are arranged in a mosaic of a
96 complex tectonic collage, amalgamated and deformed during Late Devonian through to Carboniferous Variscan
97 tectonic events (Matte et al. 1990; Aleksandrowski and Mazur 2002; Kroner et al. 2008). The collage is intruded

98 by syn- to late orogenic Carboniferous granitoid plutons, the most voluminous of which are those of Karkonosze,
99 Strzegom-Sobótka and Kłodzko - Złoty Stok (e.g. Mazur et al. 2006, 2007).

100 The basement rocks are locally overlain by syn- to late-tectonic Carboniferous intramontane basins and
101 remnants of a now much eroded post-orogenic Permo-Mesozoic sedimentary-volcanic cover (Fig. 1), which
102 during Late Cretaceous – Early Paleogene times underwent compressional deformation: gentle folding and
103 inversion-related faulting (Głuszyński and Aleksandrowski 2022). To the south, in the Intra-Sudetic Basin, Early
104 Permian red beds and shales with volcanics up to 1250 m thick are followed by Late Permian and Early Triassic
105 fluviatile deposits of Buntsandstein facies, which do not exceed a few tens of meters in thickness (e.g. Dziedzic
106 and Teisseyre 1990). To the north, in the North Sudetic Basin, there is a dominantly sedimentary series including
107 Early Permian red beds with volcanics and Late Permian Zechstein facies carbonates, sulphates and clastics,
108 averaging about 1000 m in thickness. The Early Triassic Buntsandstein of variegated sandstones, totalling about
109 600 m in thickness, are capped by Röt dolomites and later Keuper evaporites and Muschelkalk limestones (e.g.
110 Baranowski et al. 1990; Chrzastek and Wojewoda 2011). In south-western and western Poland and eastern and
111 northern Germany, the Early Triassic clastics that are accompanied locally by the Buntsandstein exceed 1000 m
112 in thickness (Feist-Burkhardt et al. 2008). The Early Permian and Early Triassic clastic deposits reflect intense
113 erosion and denudation of the southerly located Variscan internal massifs, including those of the Sudetes. The
114 Late Permian Zechstein carbonates and sulphates as well as the late-Early to Middle Triassic carbonates, on the
115 other hand, mark the episodes of marine transgression. The Late Permian transgression apparently left in the
116 Sudetes only a thin sedimentary sequence of tens of metres thick, whereas that of the late-Early to Middle
117 Triassic might have resulted in an originally much thicker sedimentary succession, similar to those in the
118 adjacent areas (cf. e.g. Feist-Burkhardt et al. 2008). This is suggested by increasingly deepening facies of the
119 now up to 100 m-thick Muschelkalk deposits towards their erosive top (Chrzastek and Wojewoda 2011). Also,
120 the original presence of a relatively thick and extensive cover of marine Jurassic sediments in the Sudetes cannot
121 be entirely ruled out, despite their present absence possibly due to pre-Cenomanian erosion.



122

123 **Fig.1** Geological map of the Sudetes (modified after Aleksandrowski et al. 1997). (1) Szklarska Poręba Huta
 124 quarry; (2) Laski quarry; (3) Laski village; (4) Chwalisław valley; (5) Mt Kopciowa.

125 2.2 Sampling

126 Our sampling was concentrated on the reddened granitic rocks in the quarry at Szklarska Poręba Huta
 127 on the Karkonosze pluton and in several exposures in the eastern part of the Kłodzko - Złoty Stok pluton, at
 128 Laski and in the Chwawlisław valley (Fig. 1). The focus was put on large exposures of rocks with distinct signs
 129 of reddening that are typically bound to fractures. In this study we use the term fracture to any brittle
 130 discontinuities in a rock, whether or not they imply fracture-parallel or -perpendicular displacement, i.e. various
 131 types of both joints and faults. Non-reddened rocks were also sampled in order to characterise differences
 132 between the reddened and non-reddened facies.

133 2.2.1 Karkonosze granitoid massif

134 The late-orogenic Karkonosze granitoid massif in the western part of the Sudetes is composed of a
 135 number of granitic facies thought to correspond to fractional crystallization, mixing with mafic magmas and
 136 hybridization processes (Žák and Klomínský 2007; Słaby and Martin 2008). The pluton resulted from sequential
 137 emplacement over a brief period of time at ca. 320-310 Ma (Machowiak and Armstrong 2007; Žák et al. 2013;
 138 Kryza et al. 2012, 2014; Kusiak et al. 2014).

139 Sampling of the Karkonosze pluton was undertaken in the Szklarska Poręba Huta quarry (N50°49'39'',
 140 E15°29'38'') on the NW margin of the Karkonosze massif, close to its contact with the metamorphic envelope
 141 (Fig. 1). Samples were obtained from a steep northern wall of the quarry where there was an exposure of two
 142 rock types in sharp intrusive contact: (1) a fine-grained equigranular leucocratic granite (so called apl granite)
 143 and (2) a medium- to coarse-grained porphyritic biotite granite. Both granites contain thick (up to 25 cm) aplite

144 veins and centimetre-wide quartz veins. According to Słaby et al. (2007), the granite from Szklarska Poręba Huta
145 belongs to the geochemically furthest evolved granite types in the Karkonosze composite pluton.

146 The porphyritic granite from the Szklarska Poręba Huta quarry contains large phenocrysts (up to 10 cm)
147 of perthitic K-feldspar (about 65-99% Or) mantled by plagioclase (about $Ab \geq 83\%$) and enclosed in medium-
148 grained groundmass (Borkowska 1966; Słaby et al. 2002). Relative to the porphyritic granite, the apl granite is
149 enriched in quartz and plagioclase and impoverished in biotite and alkali-feldspar (Kozłowski et al. 1975).

150 2.2.2 *The Kłodzko - Złoty Stok pluton*

151 The Kłodzko - Złoty Stok (KZS) pluton, the largest composite intrusion in central Sudetes (Fig. 1), is
152 composed of strongly diversified high-K metaluminous rocks, rich in mafic components (e.g. Lorenc 1994;
153 Bachliński and Bagiński 2007; Jokubauskas et al. 2014). The small magma batches were emplaced between
154 $349.0 \pm 3.7\text{Ma}$ and $331.5 \pm 2.6\text{Ma}$. Short-lived late Carboniferous rejuvenation of magmatic activity along major
155 fault zones caused intrusion of the the Laski monzogranite at $303.3 \pm 3.8/4.2\text{Ma}$ (Jokubauskas et al. 2018 and
156 references therein).

157 Exposures of reddened granitic rocks were sampled in the Laski quarry (N50°28'25" E16°48'34"),
158 Laski village (N50°27'40" E16°48'12") and a small old quarry near the creek of Mąkolnica (Chwalisław valley;
159 N50°25'54" E16°49'42"). Jointly were sampled non reddened, relatively dark-grey medium-grained granodiorite
160 with relatively uniform plagioclase composition (36-46% An) typical of the Laski area (Wierzchołowski 1976).
161 The quarry in the Chwalisław valley exposes granodiorite intruded by dykes (up to about. 1 m wide) of
162 spessartite. Both rocks had been previously dated, yielding ages of $336.7 \pm 2.3\text{Ma}$ and $321.1 \pm 3.1\text{Ma}$,
163 respectively (Mikulski et al. 2013). Non-reddened granodiorite is a grey, medium- to fine-grained, equigranular
164 rock with relative anorthite-rich plagioclase (39-44% An), whereas the fine-grained microporphyritic spessartite
165 is composed of more sodic plagioclase (Wierzchołowski 1976; Awdankiewicz and Awdankiewicz 2010;
166 Mikulski et al. 2013).

167 3 ANALYTICAL METHODS.

168 Powder X-ray diffraction (XRD) analyses were carried out at Mines-Paris, using a PANalytical X'Pert
169 system fitted with a Cu-anode and X'celerator detector, on fragments of larger K-feldspar crystals separated by
170 hand-picking from the bulk rock samples. XRD scans were run from 2 to 60° (2 θ). The XRD diagrams were
171 interpreted using the EVA© Bruker software. In the case of overlapping (131) and (1-31) peaks, deconvolution
172 was undertaken using the FITYK software (Wojdyr 2010) with Pearson VII function applied to fit the peak
173 shapes. Al-Si ordering of the K-feldspar structure was evaluated on the basis of the triclinicity parameter
174 according to the method of Goldsmith and Laves (1954).

175 Selected thin sections from the reddened rock samples were studied using conventional petrographical
176 microscopy and cathodoluminescence (CL) techniques at the Institute of Geological Sciences, University of
177 Wrocław. The CL study made use of a CL MK3 cold cathode gun mounted on a polarizing optical microscope.

178 Systematic sampling for paleomagnetic analysis was performed focused on the reddened facies at four
179 principal outcrops, three of them situated in the Kłodzko – Złoty Stok Massif (Laski quarry, Laski village,
180 Chwalisław valley) and one in Szklarska Poręba Huta area. The samples for paleomagnetic analyses were drilled

181 in the vicinity of principal fractures. The samples were collected as oriented drill cores using a petrol-driven
182 hand drilling machine and as block samples oriented in field by measuring dip and dip azimuth.

183 They were cut into standard 2.5 x 2.5 cm cylinders for paleomagnetic experiments. Thermal
184 demagnetization was subsequently performed on the cylinders either in the paleomagnetic laboratory of the
185 Institute du Physique du Globe de Paris using a vertical three-axial 2G cryogenic magnetometer and at the
186 Institute of Geophysics of the Polish Academy of Sciences in Warsaw using a 2G Enterprise SQUID
187 magnetometer. Stepwise demagnetization of the NRM was applied on the samples, starting at room-temperature,
188 and if possible, embracing the complete temperature range up to the hematite temperature interval around 700°C.
189 Alternatively, some selected samples were demagnetized using the alternating field technique at the Institute of
190 Geophysics of the Polish Academy of Sciences in Warsaw. The resulting data are systematically corrected for
191 the geographic position of each individual sample.

192 Further data processing was performed using either the PaleoMac5 software (Cogné 2003) or the
193 PALMAG software provided by Lewandoski et al. (1997) Characteristic paleomagnetic directions were
194 determined by a least-square fit of linear segments (Kirschvink, 1980) in orthogonal Zijderveld diagrams
195 (Zijderveld 1967; see also supplementary material 2, Fig. S1 to S3), either on equal areas or in great circle fits.
196 The PaleoMac5 and PALMAG software offer the opportunity to choose the grouped directions and to calculate
197 respective site-mean directions and paleomagnetic pole statistics (Fischer, 1953; Mc Fadden and McElhinny
198 1988). The analyses of thermal demagnetization of NRM curves and thermomagnetic experiments
199 (magnetization and saturation remanence during heating curves) allow the identification of magnetic carriers of
200 the paleomagnetic directions (for instrumental details and rock magnetic results please see also supplementary
201 material 2, Fig. S4 and S5).

202 The chemical compositions of feldspars and micas were analysed by EPMA (Electron Probe Micro
203 Analysis) at the Scientific and Technological Centers of the Universitat de Barcelona (CCiTUB), using Cameca
204 SX-51 and JEOL JXA-8230 electron microprobes, both operating at 15 kV and 1.5 nA. The CL images of the
205 samples were used to guide the placing of the analyses. The table of EPMA results is included in the
206 supplementary material 1.

207 Monazite dating was carried out using the EPMA U-Th-Pb_{total} method following the recommendations
208 included in Cocherie et al. (1998), Cocherie and Albarede (2001), and Williams et al. (2006). Monazite grains
209 suitable for EPMA analysis were found by means of BSE imaging using a Hitachi TM-1000 electron microscope
210 at the Manresa School of Engineering (Spain). U-Th-Pb_{total} EPMA analyses were undertaken at the Scientific and
211 Technological Centers of the Universitat de Barcelona (Spain) using a Jeol JXA-8230 electron microprobe
212 operating at 15 kV, 40 nA of current intensity and a beam diameter of 1 µm. An in-house reference monazite
213 from the Petaca District (Mexico), with an age of 1332 ± 5 Ma (2σ), was analyzed as a control standard jointly
214 with the monazite samples. The age of each analysed monazite was calculated using the MonaziteAge software
215 based on the works of Suzuki and Adachi (1991, 1994) and Suzuki et al. (1994) and which is included with the
216 Jeol JXA-8230 instrument. The initial acquired data were prepared for subsequent statistical age analysis by
217 eliminating the analyses which presented oxide totals higher and lower than $100 \pm 5\%$. Subsequently, age
218 outliers were discarded using a stepwise process based on box-plot graphs. The final ages of the different
219 monazite generations were determined by deconvolution of the ages histogram using the Microsoft Excel add-in

220 ISOPLOT 4.15 from the Berkeley Geochronology Center (Ludwig 2003). A total of 189 EPMA analyses were
221 made and 147 analyses (77% of the total) were retained including 18 from the Szklarska Poręba Huta quarry and
222 129 from the Chwalisław valley section. The table of monazite EPMA results is included in the Electronic
223 Supplementary material 1.

224 The oxygen isotope composition of plagioclase and K-feldspar was determined by Secondary Ion Mass
225 Spectrometry (SIMS) using a CAMECA IMS 1280HR ion microprobe at the Centre de Recherches
226 Pétrographiques et Géochimiques (CRPG), Vandœuvre-lès-Nancy, France. Samples were mounted in 25 mm
227 diameter epoxy mounts and the surface was metallographically polished. Afterwards the samples surfaces were
228 imaged under optical cathodoluminescence to differentiate textural domains within the crystals. The detailed
229 analytical conditions during the SIMS session are described in Fàbrega et al. (2017). The granite samples were
230 analysed jointly with in-house plagioclase and K-feldspar standards. The analyses were placed in homogeneous
231 luminescent domains in crack- and inclusion-free areas of the crystals. A total of 28 and 31 analyses were carried
232 out on plagioclase and K-feldspar grains, respectively. Afterwards, the craters of the SIMS analyses were
233 observed under BSE-imaging and analyzed by EPMA to check that the craters were homogeneously excavated
234 and determine the chemical composition linked to each isotopic analysis. The instrumental mass fractionation
235 (IMF) of the raw results was corrected by means of response surface methodology (RSM) following the
236 procedures and using the response surface models described in detail by Fàbrega et al. (2017). The SIMS results
237 including the instrumental setting, the raw and IMF-corrected $\delta^{18}\text{O}$ values, and the CL-images of the samples are
238 included in the Electronic Supplementary Information. All the $\delta^{18}\text{O}$ values of this paper are relative to VSMOW.

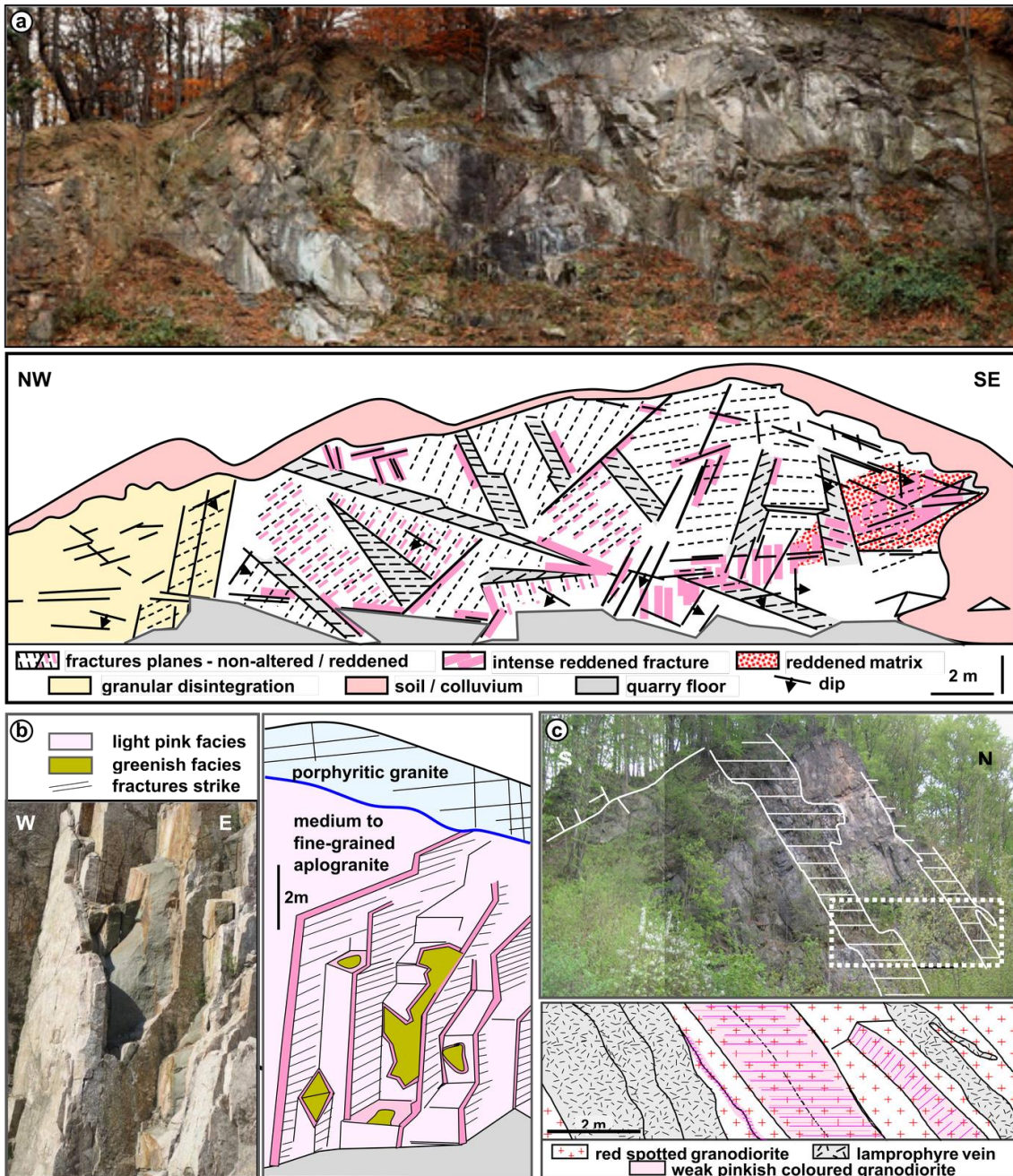
239 4 SPATIAL DISTRIBUTION OF THE REDDENING

240 A process of granite reddening has resulted in both a pervasive coloration homogeneously distributed
241 within the rock volume at some area and a more localised development of red to pinkish granite facies allied to
242 the fracture network at other areas. The most typical and widespread type of alteration is a **fracture controlled**
243 **reddening** with reddish/pinkish coloration restricted to narrow bands along fracture planes. The granodiorite of
244 the Laski quarry and the aplogranite of the Szklarska Poręba Huta quarry are typical examples for this. The most
245 intense alteration is a **pervasive and homogeneous reddening** with the entire rock impregnated by the reddish
246 pigments, and fracture planes coated with dark red iron oxides. The Laski village section is typical of this
247 pervasive reddened facies. The less intense alteration is a **red spotted facies** with pinkish spots distributed
248 within non-reddened rock. The granite of Chwalisław valley outcrop is typical of this weakly altered facies.

249 In the Laski quarry the fractures are metre-spaced, cutting the rock into well-defined 0.5 to 1 m³ blocks
250 (Fig. 2a). The extent of fracture controlled reddening is closely related to the rank of the fracture, i.e. it is weak
251 along minor fractures and more intense along major ones. The reddening penetrates the rock as deep as 1-10 cm.
252 The coloration is most intense along fracture walls but fades further away from the fractures. There is also a
253 direct relation between the frequency of the fractures and the intensity of coloration. In the SE part of the quarry,
254 the most intensive reddened facies occur along a weakly defined metre-wide zone, characterized by particularly
255 dense fracturing. Similarly, the porphyritic granodiorite exposed on Mt Kopciowa has fracture controlled
256 reddening restricted to weak, narrow staining fronts, where its large K-feldspar phenocrysts are partly or entirely
257 pink along the fractures.

258 In the Szklarska Poręba Huta quarry, the porphyritic granite is generally not reddened. However, the
259 fine to medium-grained aplogranite exhibits a light pink reddening along steep N40° major and N135° related
260 fractures, which penetrates the rock to a distance of 5-15 cm to 30-40 cm away from the fractures. Where
261 fractures are sparse, the aplogranite is greenish in the center of the blocks (Fig. 2b) but in densely fractured
262 (“chopping fractures”) zones the rock is pervasively pink.

263 The Chwalisław valley granodiorite outcrop is a typical example of the red spotted facies (Fig. 2c). A
264 weak fracture controlled pinkish coloration has developed in narrow zones along fractures but further away from
265 the fracture planes there is a scatter of spots together with small pinkish patches in the non-reddened rock. There
266 is no obvious direct relationship between lamprophyre dikes and the reddening of the granodiorite. In fact the
267 lamprophyre also has a faint pinkish coloration along the fracture planes, but their observance is hindered by the
268 dark color of the rock.



269

270 **Fig. 2** Spatial arrangement of the reddening. (a) Photo and sketch of the fracture-controlled reddening in the
 271 Laski quarry. The highly fractured zones in the SE of the quarry are the most reddened. (b) Szklarska Poręba
 272 Huta quarry facies. The reddening proceeds along the steep fractures that cut the fine-grained aplogranite. (c)
 273 Sketch of the reddish alteration facies in the Chwalisław valley outcrop. The alteration is parallel to the
 274 lamprophyre veins that intruded the granodiorite.

275 5 MICROSCOPIC ASPECTS OF REDDENING

276 Optical and cathodoluminescence petrography reveal some systematic differences between the non-
 277 reddened and reddened rock facies. We present petrographical comparison of the affected mineralogical
 278 constituents, but those that are neither affected by nor related to the staining phenomenon are omitted from the
 279 comparison.

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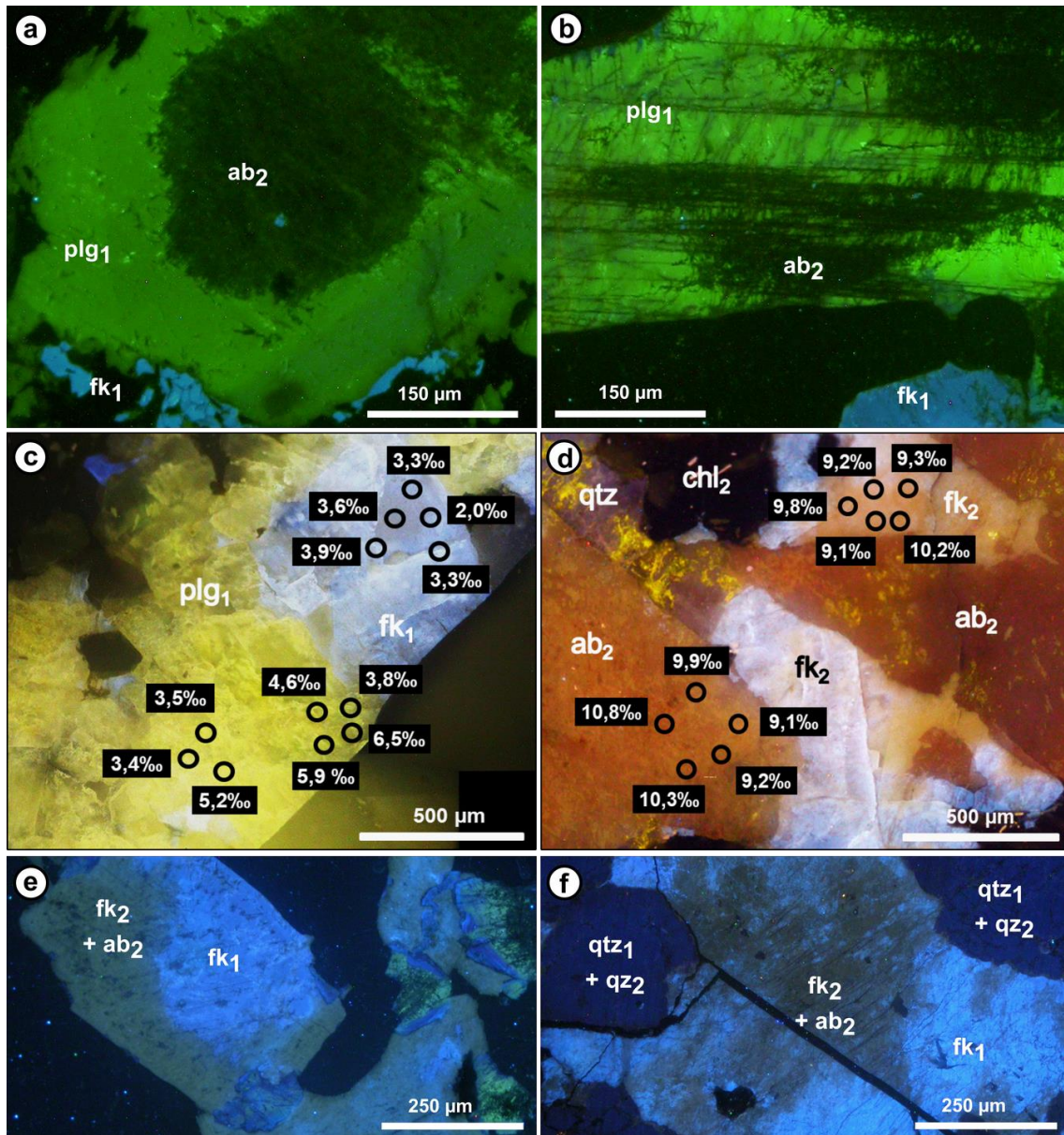
5.1 Plagioclase

In non-reddened granites plagioclase appears white to the naked eye. Large crystals are often normal-zoned with Ca-enriched cores and Ca-poor margins: some crystals are also rimmed with nearly pure albite. Polysynthetic twinning is common. It also occurs as inclusions in the K-feldspar phenocrysts (in porphyritic rock types) and constitutes most of the rock groundmass. The luminescence is typical for igneous plagioclase with yellow-green CL (Fig. 3a and 3b), that turns blue in the albitic rims. Regions with increased intensity of yellow-green CL correlates with a higher Ca content (cf. Słaby et al. 2002; Słaby and Götze 2004). Partial sericitization of primary plagioclase is often observed, especially in the Ca-enriched cores.

In reddened granites plagioclase appears pink to the naked eyes, due to hematite pigments which confer a red cloudiness to spots where primary plagioclase has been thoroughly altered to secondary untwined albite (i.e. albitization). While the hematite-free domains retain their primary CL characteristics, the pigmented plagioclase is non-luminescent (Fig. 3a and 3b), which is typical of neogenic feldspars formed at low temperatures (e.g. Marshall 1988; Götze et al. 2000; Leichmann et al. 2003; Powolny et al. 2022). Contrasting CL conveniently mark out the geometrical relationship between primary plagioclase and secondary albite: Ca-enriched crystal cores may be albitized leaving the outer Na-enriched zones unaffected (Fig. 3a) or secondary albite may be associated with microcracks, twinning and/or cleavage planes (Fig. 3b).

Primary plagioclase compositions range from An₂₆ to An₈₄, and An₁₄ to An₅₁, in the Chwalisław biotite-hornblende granodiorite and Laski valley granodiorite, respectively (Table 1). The oxygen isotope composition of the primary plagioclase, determined in the biotite-hornblende granodiorite of Chwalisław, shows $\delta^{18}\text{O}$ values ranging from +3 to +7 ‰ (Fig. 3c and Table 2), with no apparent correlation with the composition of plagioclase determined by EPMA in the SIMS points, that ranges from An₃₆ to An₄₃.

Secondary albite shows a widespread micro porosity and a typical composition about An₀₋₂, independently of the type of parent rock and of the primary plagioclase composition (Table 1). In crystals where the completeness of albitization has not been totally reached, a slightly higher Ca content was observed, with An% values ranging from 6 to 14% (Table 1). The oxygen isotope composition of the secondary albite, determined in the Chwalisław sector, is significantly heavier than the primary plagioclase, with $\delta^{18}\text{O}$ values ranging from +7 to 11‰ (Fig. 3d and Table 2).



308

309 **Fig. 3** CL Petrography and oxygen isotope composition of albitized plagioclase and K-feldspar. (a) Plagioclase
 310 with a non-luminescent albitized core and yellow luminescent rim. (b) Albitization along the major twinning
 311 planes and micro-fractures. (c) $\delta^{18}O$ values of yellow luminescent primary plagioclase and blue luminescent
 312 primary K-feldspar and (d) non-luminescent secondary albite and turbid microclinized K-feldspar. (e)
 313 Cathodoluminescence images of partially albitized K-feldspar showing non-luminescent areas containing
 314 secondary microcline, patch perthites and quartz patches on the crystal rims. (f) Albitization of K-feldspar along
 315 microfractures. plg₁: primary plagioclase; ab₂: secondary stained albite; fk₁: primary K-feldspar; fk₂: secondary
 316 K-feldspar; qtz₁: primary quartz; qtz₂: secondary quartz.

317

318 **Table 1** Mole % of feldspar components of plagioclase and K-feldspar.

site	rock	plagioclase		K-feldspar	
		primary / albitized	An %	primary / microclinized	Or %
<i>Reddened facies</i>					
Szklarska Poręba	porphyritic granite	primary albitized	1 - 3 0 - 3	microclinized	94 - 98
Laski Quarry	granodiorite	albitized	0 - 2	microclinized	95 - 98
<i>Spotted facies</i>					
Szklarska Poręba	porphyritic granite	primary albitized	8 - 14 4 - 6	partially microclinized	78 - 93
Laski Quarry	granodiorite	primary albitized	21 - 59 6 - 14	partially microclinized	79 - 92
<i>Fresh rock</i>					
Laski Valley	granodiorite	primary	14 - 51	primary	85 - 91
Chwalislaw Valley	hornblende-biotite granodiorite	primary	26 - 84	primary	72 - 93

* EPMA analyses converted to mole percent of the feldspar components by normalizing to eight oxygen atoms per formula unit.

319

320 **Table 2** $\delta^{18}\text{O}$ ‰ results of plagioclase and K-feldspar.

site	rock	plagioclase		K-feldspar primary / microclinized	$\delta^{18}\text{O}$ ‰
		primary / albitized	$\delta^{18}\text{O}$ ‰		
<i>Reddened facies</i>					
Chwalislaw Valley	hornblende-biotite granodiorite	albitized	6 - 11	microclinized	5 - 11
<i>Fresh rock</i>					
Chwalislaw Valley	hornblende-biotite granodiorite	primary	3 - 7	primary	2 - 4

321

322 Apart from hematite pigments, neogenic albite is sometimes associated with calcite and prehnite.
 323 Secondary albite is not twinned - its development erased the polysynthetic twinning in the primary plagioclase.
 324 This is in contrast to “classical” saussuritization in which the petrographic features of the primary plagioclase are
 325 preserved, at least partially (de la Roche 1957; Sandström et al. 2010; Jenkin et al. 1992; Morad et al. 2010).

326 5.2 K-feldspar

327 In non-reddened granites, K-feldspar is white to grey and often Carlsbad-twinning with varying amounts
 328 of microperthitic albite. In the porphyritic granitoids it is present as large (several centimeters) phenocrysts and
 329 as a constituent of the groundmass ($\ll 1$ cm). Cross-hatch twinning is not a common feature but, if present, is
 330 developed in domains rather than whole crystals.

331 As observed using CL, any alteration of primary crystals is manifested by a change in the blue
 332 luminescence typical of igneous K-feldspar (cf. Słaby et al. 2002; Słaby and Götze 2004; Powolny et al. 2022).
 333 The hematite-stained parts of K-feldspar crystals correspond to regions of quenched, sometimes dark-brownish
 334 luminescence, similar to the previously shown secondary albite (Fig. 3e and 3f). Such CL characteristics match
 335 secondary feldspars formed in low-temperature diagenetic and supergenic environments (e.g. Götze et al. 1999;
 336 González-Acebrón et al. 2010; González-Acebrón et al. 2012).

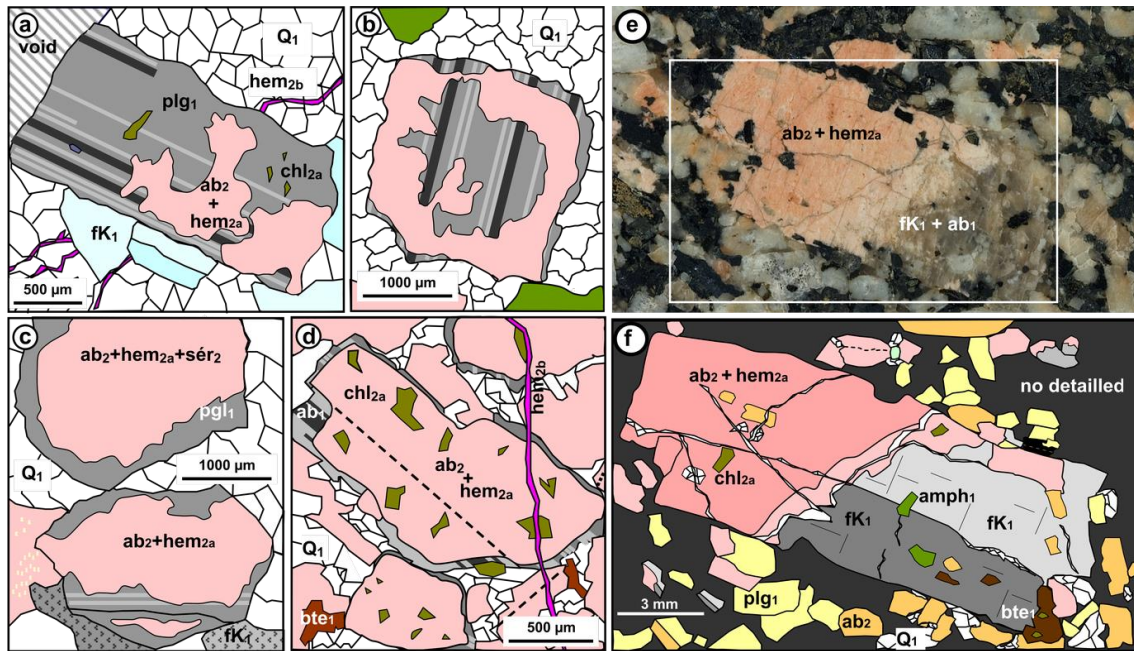
337 There is a clear relationship between hematite staining and albitization of K-feldspar. Semi-quantitative
338 measurements of the intensity of XRD reflections from the two feldspar species reveal that a low and rather
339 constant albite content in K-feldspar phenocrysts in non-reddened granitoids increases and becomes highly
340 variable in the reddened facies (Yao 2013). Primary light-colored K-feldspar phenocrysts are predominantly
341 formed of orthoclase but an increase in albite content is generally marked by its transformation (recrystallization)
342 to microcline, as indicated by an increasing triclinicity. Since triclinicity reflects the state of Al-Si ordering of K-
343 feldspar structure it is likely that the increase of Al-Si ordering, albitization and hematite-staining of the K-
344 feldspar are interrelated processes.

345 The composition of the primary K-feldspar ranged from Or₇₂ to Or₉₃ in the the Chwalisław sector and
346 from Or₈₅ to Or₉₁ in the Laski valley (Table 1). The oxygen isotope composition of the primary K-feldspar
347 (Table 2), determined in crystals from the biotite-hornblende granodiorite of Chwalislaw valley, present $\delta^{18}\text{O}$
348 values ranging from +2 to +4 ‰. The $\delta^{18}\text{O}$ values show a positive trend versus the increase in K⁺ content of the
349 K-feldspar (see Electronic Supplementary Information). In the quenched luminescence regions, the primary
350 orthoclase is replaced by secondary microcline showing abundant micro porosity, with typical composition
351 varying from Or₉₄ to Or₉₈, independently of the parent rock and the composition of the primary K-feldspar. The
352 microclinization process is accompanied by the development of patch perthites of highly pure albite, with typical
353 composition of An₀. The oxygen isotope composition of the microclinized K-feldspar (Table 2), also determined
354 in the biotite-hornblende granodiorite of Chwalislaw, presents $\delta^{18}\text{O}$ values, ranging from +5 to +11 ‰ and also
355 shows a positive trend versus the orthoclase content (Or %) of the K-feldspar.

356 **5.3 Albitization propagation**

357 Albitization of primary plagioclase occurred along microcracks and cleavage planes (Fig. 4a). The
358 albitized zones are patchy and amoeboid, with no relation to the crystallographic directions in the primary
359 plagioclase. Often albitization developed from the margins towards the central parts of a crystal, leaving the core
360 as well as the primary albitic outer rim unaffected (Fig. 4b). A common albitization pattern is observed in some
361 of the compositionally zoned plagioclases in which the Ca-enriched core has been altered but the Ca-poor rims
362 remains unaffected (Fig. 4c). The albitized core often contains scattered sericite which corresponds to an earlier
363 phenomenon than albitization as it is also observed in non-albitized plagioclase from the non reddened facies. In
364 most altered facies, primary plagioclase is pervasively albitized except outer rims composed of primary albite
365 (Fig. 4d). These altered facies are intensely reddened and often show micro-fractures filled with iron oxides.

366 **Albitization of primary K-feldspar proceeds in a similar way as in plagioclases: from the margins of the**
367 **crystals along microcracks, twinning and/or cleavage planes.** In the reddened granitoids, euhedral to sub-
368 euhedral K-feldspar phenocrysts are partly or entirely pink (Fig. 4e and 4f). There is a direct correlation between
369 the intensity of staining in the granitoids and the amount and distribution of hematite pigments in phenocrysts
370 and smaller K-feldspar crystals.



371
 372 **Fig. 4** Sketches of albitized red/pinkish plagioclase and K-feldspar. (a) Plagioclase albitization along
 373 microfractures. (b) Irregular albitization of the outer rim of the primary plagioclase: the primary albitic rim
 374 remains unaffected. (c) Albitization of the seritized core of the primary plagioclase. (d) Complete albitization of
 375 the primary plagioclase crystal except the outer rim which is composed of primary albite. (e) Polished section of
 376 a partially reddened orthoclase phenocryst. The upper part of the phenocryst is pink and dull whereas the lower
 377 part is translucent. (f) Sketch of (e). The orthoclase phenocryst exhibits classic Carlsbad twinning in the
 378 translucent part and has inclusions of hornblende, chlorite, biotite, and plagioclase. In the pinkish part the
 379 twinning is erased, and the orthoclase is albitized and hematized. fk₁: primary K-feldspar; plg₁: primary
 380 plagioclase; Q₁: primary quartz; bte₁: biotite; amph₁: amphibole; ab₂: untwined secondary albite; chl_{2a}:
 381 secondary chlorite after biotite; hem_{2a}: pigmentary hematite; hem_{2b}: coarser-crystalline hematite.

382 5.4 Fe-Mg minerals

383 Light to dark brown biotite is the main mafic mineral and minor greenish hornblende is commonly
 384 present in samples from the Kłodzko - Złoty Stok massif. In non-reddened samples both biotite and hornblende
 385 show signs of chloritization, though hornblende appears more resistant to this alteration.

386 Chloritization is significantly more intensive in the reddened granitoids where all stages of the
 387 replacement process, from initial weak alteration along the edges and cleavage planes of the primary crystals to
 388 the formation of complete chlorite pseudomorphs, could be observed. Biotite inclusions in hematite-stained
 389 plagioclase are invariably replaced by chlorite. There is no clear evidence for a volume change during the
 390 chloritization of the primary Fe-Mg minerals. Secondary microcrystalline and flaky chlorite also occurs in
 391 microfractures and intergranular boundaries. Pyrite, Ti-oxides and titanite are often associated with secondary
 392 chlorite.

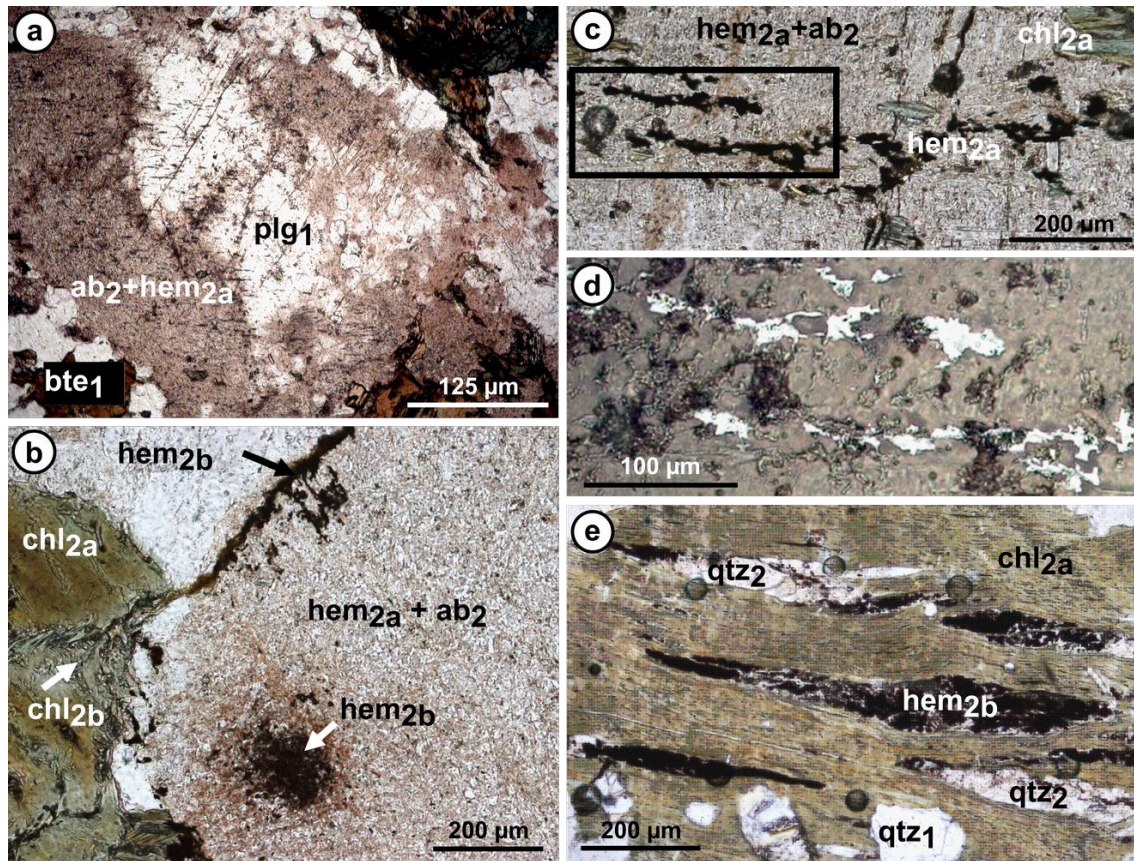
393 5.5 Fe-minerals

394 Fe-sulphides and oxides in the granitoids include pyrite, hematite, maghemite, and magnetite. Pyrite
 395 occurs irrespective of hematite pigmentation as relatively large (50-200 μm) euhedral crystals and lens-shaped

396 aggregates in the cleavages of chloritized biotite and hornblende. Late anhedral grains fill also microfractures.
397 The margins of most of the observed pyrite grains are replaced by hematite.

398 Pigmentary hematite is responsible for red cloudiness observed in plagioclase from the reddened
399 granitoid facies under microscope. In the red spotted facies plagioclase exhibits coexisting hematite-pigmented
400 and hematite-free domains which are primary igneous plagioclase (Fig. 5a) and in pervasively reddened facies
401 and in fracture controlled reddened facies the plagioclase is entirely hematite-stained (Fig. 5b). Dusty hematitic
402 pigment is confined to albitized domains in plagioclase and K-feldspar. Some observations suggest that the
403 hematitic pigments are not directly related to microporosity in the altered feldspars. Firstly, the pigment is absent
404 from the porous saussuritized feldspars of the non-reddened granitoid facies. Secondly, it is not concentrated
405 along the albitization front which is enriched in micropores that relate to the volume-loss replacement of the
406 primary feldspar by secondary albite. Instead, hematite is distributed more or less homogeneously through all of
407 the albitized feldspar. Microscopic flaky aggregates and subhedral larger grains of hematite are often observed in
408 intergranular boundaries located close to chloritized biotite and hornblende (Fig. 5b) and in microfractures
409 (Fig. 5c) and, as well as in the cleavage of some chloritized biotite (Fig. 5e), sometimes together with secondary
410 quartz and/or calcite.

411 Maghemite and magnetite could not be recognized unequivocally under the microscope nor by XRD
412 analyses but their presence was unmistakably detected by thermal demagnetization experiments (paleomagnetic
413 experiments, this study). In contrast to hematite, magnetite is restricted to the non-reddened facies. Maghemite,
414 on the other hand, was detected mainly in the weakly reddened facies. Thus, maghemite may be regarded as a
415 'transitional' Fe-mineral between the magnetite-dominated non reddened granitoids and hematite-dominated
416 reddened facies.



417

418 **Fig. 5** Hematite occurrences within albitized granitoids. a) Light pink albitized plagioclase away from reddened
 419 fracture walls with light colored relict of primary plagioclase. (b) Plagioclase entirely reddened and albitized
 420 from pervasively reddened facies; hematite appears as pigments within the secondary albite and as granules
 421 within microfractures. (c) Fine red hematite pigments and elongated hematite flakes within secondary albite;
 422 plane polarized. (d) Enlargement of (c); reflected light. (e) Hematite lenses within the cleavage of a chloritized
 423 biotite crystal, notice lenses of secondary quartz in similar position; plane polarized. plg₁: primary plagioclase;
 424 ab₂: secondary albite; bte₁: primary biotite; chl_{2a}: secondary chlorite after biotite; chl_{2b}: neogenic chlorite in
 425 micro-fracture; hem_{2a}: secondary hematite pigment; hem_{2b}: secondary granular and flaky subhedral hematite.

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5.6 Other minerals

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Quartz of the non-reddened facies is essentially unaltered, presenting a homogeneous dull-luminescence. Differently, the quartz grains of the reddened facies are significantly altered, presenting a network of micron sized non-luminescent veins of secondary quartz displaying in a cobweb texture. Internal zoning of quartz have already been described in the chloritized facies of the Sudetic Karkonosze granite (Wilamowski 2002), and similar quartz alteration has been observed in the red albitized facies of NE Spain (Fàbrega et al. 2019).

Sericite is more abundant in non-reddened samples and in weakly-reddened rock away from fractures. It occurs mostly in plagioclase and emphasizes the compositional zoning as it has preferentially developed in the Ca-enriched plagioclase cores but is entirely absent from the albitic rims. Sericitization obliterates the primary polysynthetic twinning in the plagioclase host. Tiny sericite crystals (5 μm) occur in plagioclase in the aplogranite of Szklarska Poręba Huta and less fractured granodiorites of the Kłodzko – Złoty Stok massif. This

438 form of sericite is associated with secondary albite but not with Ca-bearing minerals such as prehnite and
439 epidote. However, larger sericite flakes (10-20 μm) formed at the expense of Ca-enriched plagioclase in the
440 reddened granodiorite from Chwalisław and Laski are associated with secondary prehnite, albite, and sometimes
441 also calcite. Sericite is rare in K-feldspars.

442 Prehnite and less commonly calcite are either present as secondary minerals associated with
443 saussuritization of plagioclase and chloritization of Fe-Mg minerals or as late primary phases healing fractures.
444 While calcite is observed exclusively in the reddened rock facies, disseminated prehnite is more abundant in the
445 non-reddened granitoids. Secondary prehnite, apart from prehnite-sericite-albite aggregates in the saussuritized
446 Ca-enriched plagioclase cores, occurs also as monomineralic fan- and bowtie-shaped aggregates in intergranular
447 boundaries and as elongated lenses in biotite. The latter are concordant with cleavage and the inflation of the
448 lenses disturbs the parallel alignment of the mica flakes. They are observed only in partially chloritized biotite in
449 the non-reddened granitoids but are absent in the proximity of reddened fracture walls. Similar lenses of
450 secondary calcite are present in chloritized biotite and hornblende in the reddened rock facies. Both prehnite and
451 calcite, sometimes accompanied by laumontite, are also observed as late primary minerals healing microfractures
452 cutting across the mineral framework of the reddened granitoids.

453 Magmatic apatite is commonly present as small inclusions in biotite and hornblende or as inherited
454 inclusions in their chlorite pseudomorphs. Nonetheless, in some chlorite pseudomorphs, we also observe apatite
455 inclusions that were considerably larger and at the same time less abundant than typical apatite inclusions in
456 unaltered primary Fe-Mg minerals. They are likely a second generation of apatite resulting from recrystallization
457 and also dissolution of smaller crystals during chloritization, such as those shown in the albitized facies in the
458 Morvan (Parcerisa et al., 2009) and in NE Spain (Fàbrega et al., 2019) and those linked to low-temperature
459 spilitic albitization in the Intra-Sudetic Basin (Powolny et al. 2022).

460 REE-bearing minerals, mostly epidote and monazite, are typically subhedral to euhedral crystals around
461 50-200 μm in size, in the non-reddened. The REE-rich epidote is more abundant in the reddened facies than in
462 the unaltered rocks and occurs as larger grains up to 200-400 μm . Monazite in the reddened facies occurs in two
463 different forms depending on its position with respect to the reaction front. The crystals in a cm-scale zone inside
464 the reddish/pinkish facies in contact with the albitization reaction front are mostly pseudomorphed by secondary
465 monazite. This secondary monazite has corroded grain boundaries and micron-sized porosity, suggesting that the
466 replacement occurs via a dissolution-precipitation process. In contrast, in strongly reddened rocks, monazite
467 grains can be pseudomorphed by secondary apatite.

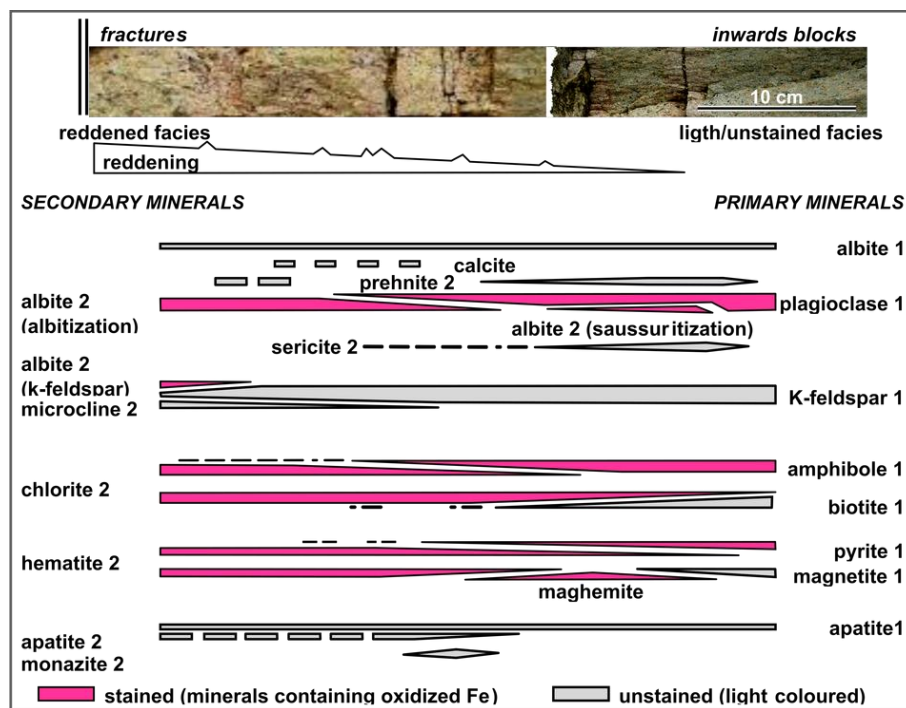
468 **5.7 Summary of petrographic observations**

469 Petrographic analyses of the reddened and non-reddened granitoids provide a basis for discerning major
470 differences in the mineralogical composition of the two granitoid facies. Apart from hematite, which is
471 responsible for the red pigmentation, and quartz, the content of which is fairly constant, the relative abundances
472 of major and some minor constituents vary with increasing reddening (Fig 6). In particular, the amount of Ca-
473 enriched plagioclase decreases largely at the expense of secondary albite and is practically undetectable in the
474 most intensely reddened sample. While plagioclase in the non-reddened rocks is partially sericitized, the
475 plagioclase in the reddened facies is predominantly albitized. To some extent K-feldspar follows the same
476 pattern, with albitization and microclinization of primary K-feldspar in the reddened facies. Weak chloritization

477 of biotite and amphibole is observed already in the non-reddened granitoids but the process is dramatically more
478 intense with increasing reddening and leads to a complete disappearance of the primary Fe-Mg minerals in the
479 most reddened facies. Chloritization of biotite, in particular where the alteration is intense, is often coupled with
480 the formation of secondary titanite and Fe-minerals along cleavage planes and microfractures. Similarly,
481 chloritized amphibole often contains calcite along cleavages, indicating the removal of Ca from the chloritized
482 crystal.

483 Small amounts of prehnite occur in the unpigmented granitoids but it becomes more abundant with
484 reddening. However, its presence also depends on the initial rock type as it is absent from the apl granite of
485 Szklarska Poręba Huta and the granodiorite of Laski village, irrespective of the extent of staining. The sericite
486 content seems to be unaffected by reddening. Dispersed calcite and laumontite are detected exclusively in the
487 reddened facies, though they are present as late fracture fillings in both stained and unstained granitoids.

488 The presence of hematite pigments in feldspars is directly related to albitized domains. For instance, we
489 have not observed albitized plagioclase that is hematite-free. Also, hematite pigments in K-feldspars were not
490 observed in non-albitized regions. The crystallization of hematite pigments seems to be synchronous with
491 albitization. If hematite formed later in micropores within secondary albite, it is expected that the migration of
492 the hematizing fluids followed some pathways. No such possible pathways were observed. Thus, assuming that
493 hematite and secondary albite are contemporaneous, the albitization must have taken place under oxidizing
494 conditions. Furthermore, secondary minerals related to albitization near the fractures are stained with iron oxides
495 whereas secondary saussurization minerals, corresponding to albite-sericite-prehnite aggregates, are not
496 affected by the reddening. Albitization-saussurization alterations grade progressively from fractures inwards
497 into the fracture blocks; they do not overlap or cut across each other as expected if they were distinct unrelated
498 alteration stages.



499

500 **Fig. 6** Spatial distribution of mineral parageneses in relation to marginal fractures and core of granitoid fracture
 501 blocks.

502 6 AGE DATING AND GEODYNAMICS

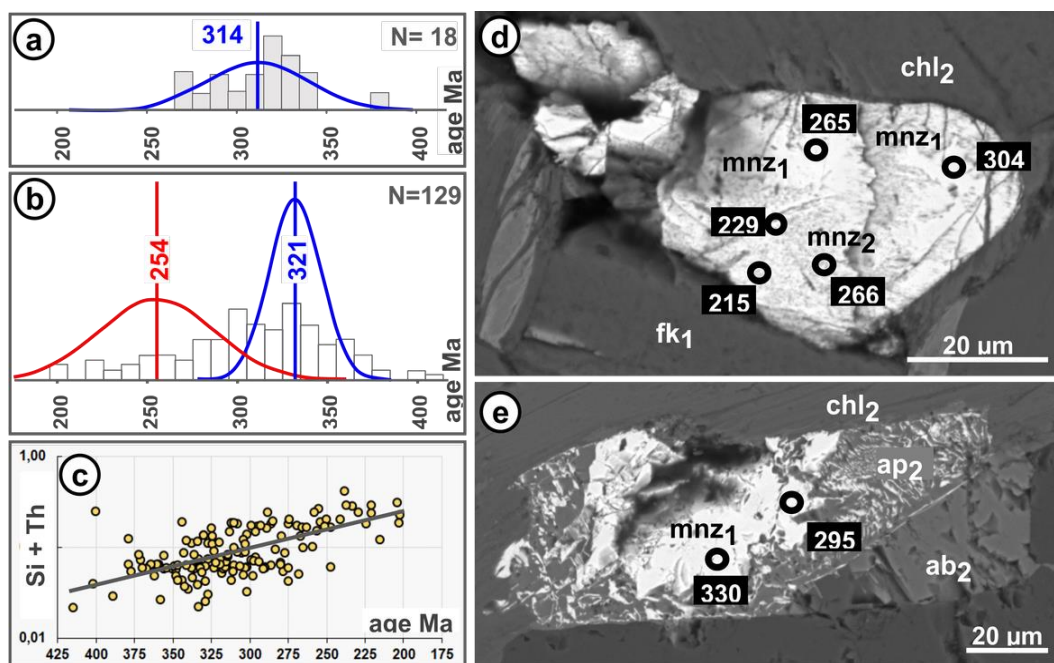
503 Dating of the albitization provides a clear tool for determining if it could be related to deep, high-
 504 temperature metasomatic processes linked to granite cooling or alternatively a product of weathering on uplift
 505 and exposure of the granitoid mass. We compared the results of two independent methods of dating: on one hand
 506 paleomagnetic dating of iron minerals and on the other hand U-Th-Pb_{total} radiochronology of monazite, a
 507 phosphate able to incorporate high U and Th concentrations.

508 6.1 Monazite dating

509 The Szklarska Poręba Huta quarry section provided only one sample containing mainly unaltered
 510 monazite. Deconvolution of the monazite U-Th-Pb_{total} ages histogram (Fig. 7a) provides an average age of
 511 314 ± 56 Ma (2σ). Although we cannot discard the possibility that some of these points could represent a mix of
 512 primary and secondary mineral ages. The Chwalisław valley section provided two samples containing numerous
 513 monazite grains for U-Th-Pb_{total} radiochronology. The deconvolution of the monazite U-Th-Pb_{total} ages histogram
 514 (Fig. 7b) yields two distinct age peaks, one at 254 ± 24 Ma (2σ) and a second at 321 ± 8 Ma (2σ) with a relative
 515 misfit of 0.988.

516 The textural analyses of samples from the Chwalisław valley section show that the older ages occur in
 517 well preserved grains from the non-reddened facies as well as in primary domains preserved within altered
 518 grains of the slightly reddened/pinkish facies. On the other hand, the younger ages (centered around 254 Ma) are
 519 almost observed in the monazite grains with dissolution-precipitation features in the reddish/pinkish facies
 520 close to the reaction front (Fig. 7d and 7e). The resetting of the monazite geochronometer is accompanied with
 521 compositional changes, including an increase in the Huttonite component [Th,Si(PO₄)], Fe and Al, and a loss of

522 REE+Y (Fig. 7c). These compositional changes in relationship with younger ages, and the systematic occurrence
 523 of the younger ages in the more altered monazite grains close to the reaction front strongly support that primary
 524 monazite in the reddish/pinkish albitized facies were variably recrystallized at about 254 Ma. The broad
 525 distribution observed in the 254 Ma peak relative to the 321 Ma peak is caused by less younger age data (in
 526 comparison with older age data).



527
 528 **Fig. 7** Dating of monazite. (a) Mean age obtained from the histogram of ages of samples from the Szklarska
 529 Poręba Huta quarry, sample WL 006. (b) Deconvolution ages obtained from the histogram of ages of samples
 530 from the Chwalisław valley section, samples WL 018 and WL 018a. (c) Single U-Th-Pb_{total} age of a monazite
 531 grain in the pinkish facies near the albitization front showing primary monazite (clear patches) partially replaced
 532 by microporous secondary monazite, BSE image. (d) Single U-Th-Pb_{total} age of a monazite grain in the strongly
 533 reddened rocks partially replaced by secondary apatite, BSE image. mnz₁: primary monazite; mnz₂: secondary
 534 monazite; ap₂: secondary apatite; ch₂: chloritized biotite; fk₁: primary K-feldspar, ab₂: secondary albite.

535 6.2 Paleomagnetic dating

536 Albitization and reddening caused by the presence of hematite, are shown to be closely associated as
 537 they reveal no spatial discrepancy at outcrop or microscope scale. We thus consider albitization and hematization
 538 to be cogenetic. The dating of the hematite formation is possible by dating the paleomagnetic secondary
 539 remagnetization signal. This would then allow to date the albitization event.

540 The different components of the paleomagnetic signal recorded within the analyzed oriented samples
 541 have been separated by stepwise thermal demagnetization experiments and are related to their specific magnetic
 542 carriers (see also supplementary material 2). According to the respective temperature interval at which a specific
 543 decay in the magnetic intensity curve was observed, one or a succession of several magnetic carriers were
 544 identified in the sample. In principle, thermomagnetic and microscopic analysis showed three different types of
 545 magnetic mineral assemblages were observed within the analyzed samples: (1) hematite dominated in pervasive
 546 reddened facies and in fracture controlled reddened facies; (2) coexisting hematite-maghemite mostly in fracture

547 controlled reddened facies further away from the fractures; and (3) maghemite dominated mainly in spotted
548 facies.

549 In addition, maghemite-bearing mineral assemblages sometimes register residual, presumably primary
550 magnetite. This was mainly observed in samples distant from fracture zones and in the less reddened spotted
551 facies. As well, a weak contribution of goethite was sometimes observed which was demagnetized after
552 application the 120°C demagnetization step. Goethite is considered as negligible component compared to the
553 other magnetic carrier minerals and is most likely linked to a weak recent overprint by surface weathering of the
554 outcropping rock sections.

555 The stepwise thermal demagnetization allowed us to obtain the vector directions recorded by the
556 magnetic carrier minerals at the moment of their formation. These directions were considered to be characteristic
557 for either the maghemite or hematite components, or both, if their inclination and declination values were stable
558 according to the respective considered temperature intervals. It is thus possible to sometimes extract two slightly
559 different directional intervals from one single sample if the two main carriers diverge in their mean direction.
560 The results were 'sub-grouped' according to their directional information.

561 The extracted directional information was then transformed into magnetic paleopoles (e.g. Butler 1998)
562 that were subsequently projected onto the apparent polar wander path (APWP) reference curve for Eurasia (Edel
563 and Düringer 1997; Besse and Courtillot 2003). In total 275 individual samples were analyzed for the calculation
564 of the mean paleopoles carried by hematite or hematite-maghemite.

565 The severe rejection criteria for paleomagnetic samples in our dataset are mainly due to high error
566 defined by values of α_{95} (Kirschvink 1980; Fisher et al. 1987). Additionally, we eliminated samples that did
567 not show any linear segments of the demagnetization vector on a Zijderveld diagram or a clear stable endpoint
568 near the origin of a Zijderveld plot, as described by McFadden and McElhinny (1988). The resulting 18 averaged
569 paleopoles all show acceptable α_{95} values (< 15) according to the recommendations of e.g. Fisher et al. (1987)
570 and very often plot directly on or close to the Apparent Polar Wander Path (APWP; Table 3). Six paleopoles
571 were calculated for one or two directions only. For these paleopoles the mean angle deviation (MAD) values are
572 given instead. All poles were listed in the Table 3. The paleomagnetic data set also includes additional
573 paleopoles that correspond to the primary Devonian magnetization of the granitoid rocks, carried by primary
574 magnetite (not shown here).

575

576 **Table 3** Results from paleomagnetic data analyses, corresponding to the averaged site-wise paleopoles plotted in
577 Fig. 8. MAD = mean angle deviation; TH = thermal demagnetization; AF = alternating field demagnetization;
578 mgt = magnetite; mhm = maghemite; hm = hematite; goe = goethite; LC = low coercivity compounds;
579 HC = high coercivity compounds.

paleo- poles	site outcrop	age	Nb of samples	α 95	MAD	type of analysis	pole latitude	pole longitude	magnetic carriers
1	Sklaska Poreba	Late Carboniferous	3	8.9	10.1	TH	20	125	mgt, mhm, hm
2	Chwalisław	Late Carboniferous	2			AF	23	126	LC+HC
3	Laski quarry	Late Carboniferous	3	8.7		TH	27	141	mgt, mhm, hm
4	Laski quarry	Middle Permian	4	8.7	6.7	TH	41	182	mhm, hm
5	Laski quarry	Late Permian	2			TH	53	161	goe*, hm
6	Laski village	Late Permian	1			TH	55	170	hm
7	Sklaska Poreba	Late Permian	3	10.1	7.7	TH	47	157	mhm
8	Laski quarry	Late Triassic	4	5.1		AF	54	127	LC+HC
9	Laski quarry	Triassic/Jurassic	2			AF	51	107	HC
10	Chwalisław	Triassic/Jurassic	3	3.3	6.6	AF + TH	49	118	mhm
11	Chwalisław	Triassic/Jurassic	3	7.3		TH	42	99	mhm, hm, goe*
12	Sklaska Poreba	Triassic/Jurassic	3	7.2		TH	36	104	mhm, hm
13	Laski quarry	Early Jurassic	4	9.1	6.6	TH	53	74	mhm
14	Chwalisław	Early Jurassic	8	5.9		TH	58	83	mhm, goe*
15	Laski quarry	Early Jurassic	2			AF + TH	73	93	mhm, hm
16	Sklaska Poreba	Early Jurassic	3	5.9	10.4	AF	74	67	LC+HC
17	Laski village	Middle Jurassic	6	5.0		TH	73	170	hm
18	Laski village	Middle Jurassic	1			TH	79	114	hm

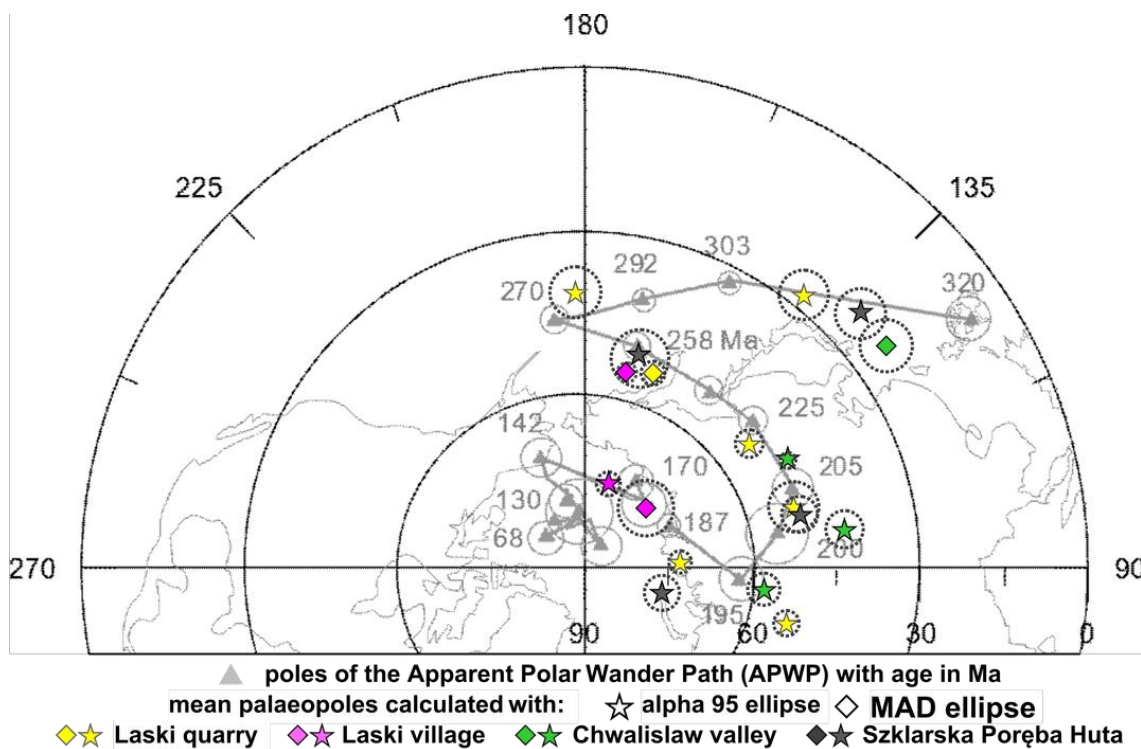
580

581 Figure 8 shows the mean paleopoles for all four studied outcrops including Laski quarry, Laski village,
582 Chwalisław valley, and Szklarska Poręba Huta. Four main groups of paleopoles and more 'isolated' average
583 paleopoles plot along the APWP curve for Eurasia:

- 584 (1) Samples from paleopoles 1, 2, and 3 have a mixed magnetic assemblage which is dominated by primary
585 magnetite. These represent the oldest group with Late Carboniferous ages (between 320 to 303 Ma) and are
586 represented in all outcrops (with exception of Laski Village, that exhibits exclusively pervasive reddened
587 facies) whichpresumably correspond to the primary age of the granitoid. All other paleopoles record
588 secondary remagnetization ages.
- 589 (2) An 'isolated' average paleopole of Middle Permian age near the 270 Ma reference point from Laski quarry
590 (paleopole 4).
- 591 (3) A cluster of Late Permian ages close to 258 Ma from the Szklarska Poręba Huta quarry, Laski quarry and
592 the basal part of the Laski village section (paleopoles 5, 6, 7).
- 593 (4) A relatively 'isolated' average paleopole of Late Triassic age close to 225 Ma from Laski quarry
594 (paleopole 8).
- 595 (5) A group of ages near Triassic/Jurassic boundary (around 205 Ma) comprising four sets of paleopoles from
596 the Laski quarry (paleopole 9), Chwalisław valley (paleopoles 10 and 11) and Szklarska Poręba Huta
597 (paleopole 12) samples.
- 598 (6) A group of paleopoles of Middle Jurassic ages (between 175-165 Ma; paleopoles 17 and 18). This group is
599 clearly different because it is solely carried by samples from the upper part of the Laski village section that
600 is pervasively reddened/hematized. The intense remagnetization of these samples seems to have erased the

601 primary magnetization carried by magnetite and the potential Permo-Triassic overprint. The lower part of
 602 the section still retains the Permo-Triassic signal.

603 (7) A group of Early Jurassic paleopoles (13, 14, 15, 16) plots between the 195 to 187 Ma reference point, but
 604 they show a significant S to SW offset from the APWP. Part of this directional offset may be due to the fact
 605 that these samples (from the Laski quarry, Chwalisław valley, and Szklarska Poręba Huta) may have been
 606 remagnetized successively multiple times during the Permo-Triassic and later again during Early Jurassic
 607 times. If so, the hematite in these samples could carry a mixed cumulative age signal that is probably less
 608 well extractable than the initial remagnetization from primary magnetite to maghemite and/or hematite
 609 during the Permo-Triassic alteration phase.



611 **Fig. 8** Stereographic projection of the paleomagnetic results for the calculated average paleopoles for the study
 612 sites. The paleopoles are compared to the APWP for Eurasia (grey line) using the combined curve of Edel and
 613 Düringer (1997) from the Permian to Middle Jurassic (grey circles) and Besse and Courtillot (2003) from the
 614 Middle Jurassic until present (grey triangles). Star symbols represent mean poles with their respective α 95
 615 confidence circle, diamond symbols are paleopoles with their respective MAD confidence circle, corresponding
 616 to Table 3.

617 In our petrographic studies magnetite was solely found in the non-reddened facies whereas the
 618 magnetite detected by thermal demagnetization in samples of the weakly reddened facies probably relates to
 619 residual tiny crystals included in quartz. Maghemite is mainly detected in the weakly to intermediately
 620 pigmented facies distant from the reddened fractures at Laski and Szklarska Poręba Huta and in red spotted
 621 facies from Chwalisław valley. According to the literature, maghemite is metastable with respect to hematite and
 622 is considered to have formed by the slow oxidation of magnetite under weak oxidative conditions (Steinthorsson
 623 and Helgason 1992; Bowles et al. 2011). Hematite is dominant in the pervasively reddened facies in the Laski
 624 village section and is indicative of strong oxidative environments.

625 A chronology appears between the different magnetic facies of alteration. The less reddened spotted
626 facies which is mainly dominated by maghemite carries the oldest remagnetization ages while the pervasively
627 reddened and albitized facies shows the youngest ages. In the mean time, incipient reddening and albitization
628 appears as the less intensely altered facies and conversely the pervasively reddened and even pervasively
629 albitized without primary plagioclase and K-feldspar relict appears as the most intensely altered facies. This
630 conjunction and the fracture controlled reddening that diminish obviously from most intensely altered facies
631 towards the less altered spotted facies led us to consider that these various gradients agree with a succession of
632 facies from the base to the top of the alteration profiles. This suggests also that, with time, the active alteration
633 front progressively retreated upwards into shallower weathering horizons.

634 It is of particular note that the **Late Permian** U-Th-Pb_{total} age at 254 ± 24 Ma obtained for secondary
635 monazites in samples from the Chwalisław valley section corresponds to the paleomagnetic poles **10, 11, and 14**
636 near 258 Ma on the Eurasian APWP (Fig. 8). These paleomagnetic poles, were obtained for weakly pigmented
637 facies from Lasky quarry, Sklarska Proeba Huta, and Laski village, from samples taken further away from
638 reddened fractures that contain significant amounts of maghemite. On one hand, maghemite developed in the
639 early stages of albitization, under less oxidative conditions. On the other hand, secondary monazite formed in the
640 immediate vicinity of the albitization front and was rapidly altered away from this front. The consistency in
641 independent ages of the albitizing and oxidizing environments obtained from completely different secondary
642 minerals and by means of distinctly different dating methods provide confidence in datation results. This does
643 not contradict with the younger paleomagnetic ages obtained in this study for which we do not have access to
644 separately datable minerals.

645 **6.3 Links to other age determinations**

646 The older age obtained for monazite from the Chwalisław valley section (321 ± 8 Ma) is within the range
647 of the previously published primary crystallization ages for the Kłodzko - Złoty Stok (KZS) Pluton between
648 349.0 ± 3.7 Ma and 331.5 ± 2.6 Ma (U-Pb, zircon; Jakubauskas et al. 2018 and references therein). The younger
649 age (254 ± 24 Ma) reflects growth of secondary monazite via dissolution-reprecipitation process. Similar Permian
650 K-Ar ages between 256-268 Ma (K-Ar isochron age, 4 points: 261 Ma) were reported by Depciuch (1972) for
651 KZS diorites from Małolno affected by chloritization/sericitization and interpreted as time of hydrothermal
652 solutions activity. Unaltered samples representing main varieties of KZS igneous rocks provided significantly
653 older K-Ar ages ranging from 291 to 303 Ma (K-Ar isochron age, 14 points: 295 Ma).

654 Our mean value of the monazite age (314 Ma) for Szklarska Poręba Huta aplogranite is consistent with
655 high precision ID-TIMS U-Pb zircon ages 312.5 ± 0.3 Ma and 312.2 ± 0.3 Ma indicating crystallization of two
656 main granite varieties (porphyritic and equigranular) of the Karkonosze Pluton (Kryza et al. 2014). Similar
657 $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 312 ± 4 Ma reported by Marheine et al. (2002) suggest fast cooling below closure
658 temperature of the K-Ar system. It has to be mentioned, even if of poor quality, an age of 271 ± 20 Ma obtained
659 on monazite from the same aplogranite at Szklarska Poręba Huta (Mikulski et al. 2004). This dating deserves to
660 be linked to the 254 ± 24 Ma obtained on the secondary monazite of the Chwalisław valley section. The
661 Karkonosze granite underwent fast uplift to the surface as evidenced by granitic detrital material in Upper
662 Carboniferous sediments of the North Sudetic Basin (Milewicz 1968, Milewicz 1970).

663 Previous paleomagnetic dating aimed to determine the primary age of the emplacement of several
664 Carboniferous plutonic units of the Polish Sudetes and led to the detection of Permo-Triassic (re)magnetizations
665 (Edel et al. 1997). Even if the study did not plan to determine the age of the albitization, and even if sampling of
666 the obviously altered (reddened) facies was avoided, re-magnetization events, younger than the age of the
667 emplacement of the granitoid, were present in some of the samples from the Chwalisław valley amphibolites in
668 contact with the Kłodzko - Złoty Stok intrusion and in the Izera dolerite close to the Szklarska Poręba Huta
669 quarry (Edel et al. 1997). The authors identified paleomagnetic carriers of these Permo-Triassic paleopoles as
670 hematite and maghemite, while the Carboniferous paleopoles reported by Edel et al. (1997) are carried by
671 (primary) magnetite. Further paleomagnetic studies of Kądziałko-Hofmokl et al. (2003) on the Kłodzko
672 Metamorphic Complex report on Triassic and Jurassic directions for similar magnetic mineral assemblages
673 (Kądziałko-Hofmokl et al. 2003).

674 The Permian-Triassic age in the Kłodzko – Złoty Stok massif and the Szklarska Poręba area recurs
675 systematically despite the different dating techniques. Thus, there is consistency between the results of
676 paleomagnetic dating of the reddened/albitized facies in this study and other published age determinations on the
677 various facies which were simply described as ‘altered’ but not clearly characterized as albitized.

678 **6.4 Link to the Post-Variscan paleosurface**

679 The Sudetic area, together with the southern part of the Germanic Basin of central Europe, is ascribed to
680 have been emergent during Permian and Trias on the basis of stratigraphic and sedimentological evidence (Feist-
681 Burkhardt et al. 2008; Głuszyński and Aleksandrowski 2022). At the same time, Permian and Triassic ZHe
682 cooling ages reported by Danišik et al. (2010) and Sobczyk et al. (2015) are consistent with the idea of exposure
683 of some parts of the Karkonosze granite during the Triassic (see also Migoń and Danišik 2012). Similarly,
684 Carboniferous to Triassic ZHe data reported by Danišik et al. (2012) show that certain parts of the Kłodzko -
685 Złoty Stok granitoid massif may have been close to the Earth surface during the Triassic ($249.8 \pm 20.0 -$
686 209.4 ± 4.1 Ma) and reflect the record of progressive cooling through the zircon helium partial retention zone.

687 Results of zircon (U-Th)/He (ZHe) dating available for the Western Sudetes area revealed i.a. Triassic
688 ($248.2 \pm 16.6 - 212 \pm 14.2$ Ma) cooling signals (Sobczyk et al. 2015). They have been interpreted as a record of
689 bedrock slow exhumation or alternatively as an apparent ages due to zircon thermal system rejuvenation, as a
690 result of burial (see discussion in Sobczyk et al. 2020) or an advective heat flow under crustal extensional
691 conditions (Botor et al. 2017). Noteworthy, ZHe datings for the Permian sandstones from the Fore-Sudetic
692 Homocline (Sobczyk et al. 2013) provided a single Mid-Triassic cooling age (244 ± 5.0 Ma) within a
693 Weissligendes series, correlating with paleomagnetic datation which was interpreted, due to its clear post-
694 depositional origin, as a probable age of Kupferschiefer mineralization. Therefore, low-temperature
695 thermochronological data, even if not providing indisputable proof for the Post-Variscan paleosurface exposure
696 during Trias, give some arguments for linking Triassic bedrock cooling with a regional crustal instability during
697 the post-orogenic unroofing. Also the local sedimentological record for the Early Triassic lithofacies within the
698 Intra-Sudetic and North Sudetic synclinoriums indicates channel flow and river bed depositional environments
699 (Kowalski 2020), suggesting the predominance of continental conditions for this period with local source areas
700 located mainly in the Karkonosze-Izera Massif. Noteworthy, strongly kaolonized undisturbed Triassic
701 (Buntsandstein) sandstones from the Sudetes are interpreted as a record of *in situ* chemical weathering profiles at

702 the Triassic paleosurface (Kowalski 2020 and references therein). Finally, regional review studies indicate that
703 during the Late Triassic to Early Cretaceous the Sudetes landforms were steady sustained by the Post-Variscan
704 paleosurface that formed a vast land mass undergoing slow erosion and denudation resulting i.a. in formation of
705 deep weathering mantles, known also from central and northern Europe (Migoń and Lidmar-Bergström 2001).
706 Powolny et al. (2022) advocated a late to post-magmatic low-temperature fluid-rock interactions controlling the
707 process of albitization within the Permian trachyandesites from the Intra-Sudetic Basin. Based on apatite fission-
708 track central ages obtained for three samples, these authors proposed a Jurassic (182 – 161 Ma) time interval for
709 this thermal event. Closer inspection of apatites analyzed by Powolny et al. (2022) show that >50% of all dated
710 grains revealed a Triassic cooling age, which indicates a possible two stage post-Variscan albitization in the
711 Sudetes.

712 In the absence of any major Triassic magmatic event in the Sudetes and adjacent areas, we link the
713 reddening alteration of the Kłodzko – Złoty Stok and of the Karkonosze granites, as well as their albitization, to
714 a Post-Variscan/Pre-Triassic paleosurface, similar as described for the French Massif Central (e.g. Schmitt and
715 Clement 1989; Parcerisa et al. 2009) and in NE Spain (Fàbrega et al. 2019). It follows then that the most
716 pervasively reddened/albitized facies were formed near the paleosurface while at depth the granites were
717 progressively less affected by the alteration process. It is of note that the conspicuous reddened and albitized
718 fractures are never cut by any fractures that do not exhibit this alteration. Hence, we regard the lack of non-
719 albitized fractures as evidence that the alteration is relatively recent and post-dates the exhumation of the granite
720 bodies.

721 Moreover, the superficial origin of albitisation and reddening via weathering explains the spread of the
722 paleomagnetic data along the APWP curve for 100 Ma from the Permian to the Jurassic. This distribution of
723 palaeomagnetic ages is not the result of successive phases of alteration, but rather a single event that is
724 heterogeneously recorded in the rock. It relates to the progression of weathering along fractures and even around
725 kilometer sized massive rock volumes. But presumably there were times of more active and less active phases
726 during this long period, mainly under the influence of climate and eustatic sea level changes, which could
727 explain some clustering of paleomagnetic ages during distinct periods.

728 **7 INTERPRETATION OF THE PARAGENESES**

729 In general, our observations agree with earlier data on fracture related hydrothermal alterations with the
730 development of albitization and saussuritization (e.g. De Jong and Williams 1995; Lee and Parsons 1997;
731 Engvik et al. 2008; Plümper and Putnis 2009; Sandström et al. 2010; Morad et al. 2010; Powolny et al. 2022).
732 The direct relationship of albitization and related alterations with the Post-Variscan paleosurface completely
733 changes the paleoenvironmental conditions that prevailed for these alterations compared to quoted studies that
734 ascribe them a deep and relatively high temperature environment. Nevertheless, in the context of a direct
735 relationship with the Post-Variscan paleosurface, the alterations discussed in this study do not exhibit any
736 typical regolith features, such as illuviation structures related to water percolation in vadose environments,
737 bioturbation, or pallid or iron stained mottles. On the contrary, the isotropic structures around blocks evoke an
738 aquifer environment and recall saprolite from the lower zones of regolith horizons corresponding to deep
739 weathering of bedrock. The most astonishing characteristic is the alteration around the fracture blocks. This

740 occurs without modification of the primary fabric of the rock and without any obvious development of porosity.
741 However, the alteration developed systematically via a hierarchical net of fracturing from the macro-scale,
742 demarking fracture blocks, to intergrain microcracks, giving to the albitization front at outcrop scale a dendritic
743 aspect of fresh unaltered blocks surrounded by reddened fractures (Fig. 9a). Here we examine the geochemical
744 grades, coupled dissolutions, precipitations and water-rock interactions in order to identify the nature of the
745 fluids that triggered these alterations and ultimately the nature of the related paleoenvironment.

746 **7.1 Fracture-bound geochemical grades**

747 The alteration primarily and intensively affects the plagioclase and only to a lesser extent K-feldspar
748 and ferromagnesian minerals (biotite and amphibole). The alteration of plagioclase is of three kinds (Fig. 9a and
749 16b).

- 750 1) Albitization of the primary feldspars is causally linked to proximity to the fractures. Some late Ca-bearing
751 silicates may be associated with this alteration, but they rather fill in microcracks and their origin may not be
752 directly related to the albitized plagioclases.
- 753 2) Saussuritization of the primary plagioclases accompanied by the development of Ca-bearing prehnite is
754 typical of the non reddened and lightly reddened facies away from fractures. Secondary albite has developed
755 along with saussuritization of the primary plagioclase.
- 756 3) Invasive development of sericite (sericitization) affects in places the primary plagioclases from the non
757 reddened and lightly reddened facies and is associated with the formation of small neogenic albite but
758 without any Ca-bearing silicates.

759 The alteration of ferromagnesian minerals is connected to the alteration of plagioclase. There is a
760 progressive chloritization of biotite in the non reddened and lightly reddened facies as the fractures are
761 approached. Chloritization seems to occur either under volume conservation or with a slight decrease of the
762 volume of the biotite with secondary minerals such as quartz, K-feldspar, prehnite, and iron-bearing minerals
763 embedded in the cleavage planes. The amphiboles appear to be less sensitive to chloritization than biotite,
764 remaining practically unaltered in the unaltered and lightly reddened saussuritized facies.

765 These alterations result in a strong geochemical overprint between the core and the boundaries of the
766 fracture-isolated blocks (Fig. 9c). The margins of the blocks are strongly Na-enriched due to albitization of
767 plagioclase, and depleted in K due to chloritization of biotite and partial albitization of K-feldspar. The cores of
768 the blocks are relatively enriched in K, Ca and Al compared to the albitized facies (taking into consideration the
769 saussuritization of the plagioclases).

770 The secondary mineral parageneses, especially albitization and saussuritization, never exhibit an
771 interruption or clear alteration front but appear rather as continuum, grading from one to another. There is no
772 obvious temporal succession between the parageneses. By contrast, the parageneses are overlapping, with small
773 sericites in the albitized/hematitized plagioclase crystal cores formed due to alteration of the plagioclase before
774 albitization. The secondary parageneses appear more like an expression of the geochemical gradients between
775 fluids circulating within the fractures and the massive granitoid blocks. Thus, the observed sequences of
776 neogenic minerals result from the progressive encroachment of the reaction front towards the cores of the
777 fracture blocks while their outer margins are gradually altered in equilibrium with the interstitial fluids. It is

778 appropriate at this juncture to refer to metasomatic sequence or zonation rather than paragenesis involving
779 successive or different alteration processes.

780 The albitization as well as the iron oxide pigmentation becomes less intense away from the walls of the
781 fractures towards the inner parts of the fracture blocks. This gradient reflects a decreasing Na^+ and O_2 (and/or
782 Fe) content of the fluids circulating in the intergranular crystal boundaries (Fig. 9d). In contrast, the K^+ and Ca^{++}
783 content of the interstitial fluids would have decreased in the opposite direction: K-bearing minerals (K-feldspar,
784 biotites) and Ca-bearing plagioclases were altered by leaching of K^+ and Ca^{++} at the fracture walls and sericite
785 formed within the lightly reddened and unaltered facies progressively more distant from them.

786 **7.2 Coupled dissolution-precipitation reactions**

787 It is well known, plagioclase with a high An content is susceptible to preferential dissolution due to
788 enhanced instability in the presence of water (Matthews and Goldsmith 1984; Stillings and Brandtley 1995).
789 Thus, destabilization of plagioclase could be a trigger resulting in a chain of reactions in the altered granitoids.

790 Albitization of plagioclase proceeds at constant volume (with low porosity) given a Na-supply,
791 additional silica, and with the release of Ca and some Al (Ramseyer et al. 1992; Parcerisa et al. 2009;
792 Hövelmann et al. 2010). In the vicinity of the fractures, the released cations would have been leached and
793 evacuated by the circulating fluids. As the reaction continued, fluid circulating in the rock by diffusion and
794 convection would have become gradually depleted in Na, and albitization diminished and or even ceased at low
795 Na-content (Fig. 9d).

796 Biotites are destabilized and transformed into chlorite in the vicinity of the fractures, probably through a
797 double effect of the leaching of the K interlayer cations and oxidation of the Fe^{2+} in octahedral sheets (Fig. 9e).
798 Here leaching of $\text{Si}(\text{OH})_4$ takes place because the cation/Si ratio is equal to ~1 within the biotite, whereas it is
799 between 1 and 1.5 within the replacement chlorite. Detailed study of the Karkonosze granite's chloritization
800 (Wilamowski 2002) showed that "The rate of potassium removal from biotite during the progress of
801 chloritization is not constant. The rapid removal of K is observed at initial stages of chloritization. After the
802 replacement of 20% of the biotite, K removal is much slower." The released silica could have nourished the
803 albitization of the plagioclase. If the chloritization occurred without an additional cation supply, the volume loss
804 would have been about 30%, whereas if additional cations were introduced (such as the Al released by
805 plagioclase alteration) the volume loss may have been less substantial (Ferry 1979).

806 Amphiboles appear to be less "sensitive" to chloritization than the biotites, probably because their
807 structure does not contain any unstable or mobile cations such as K^+ . The destabilization of the amphiboles is
808 most likely a result of leaching of Mg and perhaps oxidation of Fe^{2+} . It might be that amphiboles, and in
809 particular hornblendes (which are calcic), are relatively protected against chloritization because of the Ca
810 liberated by albitization of the plagioclase. For that reason the chloritization of amphiboles occurs only in the
811 more intensively altered facies in which plagioclases are albitized and therefore no longer provide Ca^{++} to the
812 fluids.

813 Hematization is systematically associated with the albitization. Hematite occurs mainly as pigments
814 within the neogenic albite or as coarser crystalline material associated with the ferromagnesian minerals. There
815 is no iron oxide precipitation in the fractures, so the fluids circulating in the fractures appear unlikely to have

816 been iron-bearing. Hematite has likely been nourished by the iron released by alteration of ferromagnesian
817 minerals.

818 The formation of sericite depended mainly on the availability of K^+ . Chloritization of biotite was the
819 quasi-exclusive source of K in the system because albitization of K-feldspar was not pervasive. Besides, K
820 released by chloritization in the early stage of the alteration and in the vicinity of fractures would have been
821 evacuated by convective fluid flow. Sericitization is present away from the fractures where diffusion was
822 dominant with respect to the movement of the chemical elements. Thus, sericitization is dominant in the non
823 reddened and moderately reddened facies where the Na-content was lowered in the interstitial fluids due the
824 increasing distance from the fractures that are the major source of Na. Yet, where the K/Na ratio was higher,
825 albitization was limited and K competed with Na to form K-bearing silicates (mainly secondary K-feldspars
826 and/or sericite) in combination with Si and Al (Ben Baccar et al. 1993; Perez and Boles 2005).

827 Euhedral crystals of prehnite within altered plagioclase indicate epigenetic replacement with the main
828 chemical component directly supplied in situ by the plagioclase. On the other hand, prehnite finning in
829 microfractures witnesses the mobility of these components over some distance. If the mobility of Ca and Si is to
830 be expected, it is rather unusual for Al, typical fairly immobile in a medium close to neutral pH, except in
831 presence of complexing agents (Castet et al. 1993; Farmer and Lumsdon 1994; Oelkers et al. 1994). In any case,
832 Al appears to have been fairly mobile, at least in some cases, but it is difficult to assess whether or not Al was
833 evacuated from the fractures by fluid circulation.

834 Recrystallization of monazite was also apparently triggered by calcium availability. Initially, primary
835 monazite recrystallized into secondary monazite at the early stages of albitization but later, during advanced
836 stages of the albitization, primary and secondary monazite recrystallized into apatite in concert with the increase
837 in calcium availability in the system.

838 These alterations can be considered to be topochemical: the saussuritization of plagioclase provided Ca
839 for prehnite development and the chloritization of biotite is the likely source of K for sericitization. However, the
840 complete albitization of the plagioclases and in particular the albitization of the K-feldspars in the vicinity of the
841 fractures testify for the supply of Na, and the hematite associated with albite registers the oxidative capacity of
842 the alteration and therefore a supply of oxygen.

843 **7.3 Fluid-rock interactions**

844 The geochemical exchanges between reaction zones are determined by the transport of the ions in the
845 fluids. Exchanges in fractures record the circulation of the transport fluids by convection. The circulation may
846 have been relatively rapid in the major fractures but slower as the dimensions of the fractures decreased. Away
847 from the open micro-fractures, the quantity of fluid is reduced and mainly limited to 'thin films' at the
848 boundaries of mineral grains and the dissolution-recrystallization interface. Such confined fluids may be very
849 different in chemical composition and behavior from bulk fluids (Putnis and Ruiz-Agudo 2013). Within thin
850 films, the circulation would have been minimal, diffusion prevailed over convection and ion mobility was
851 extremely reduced.

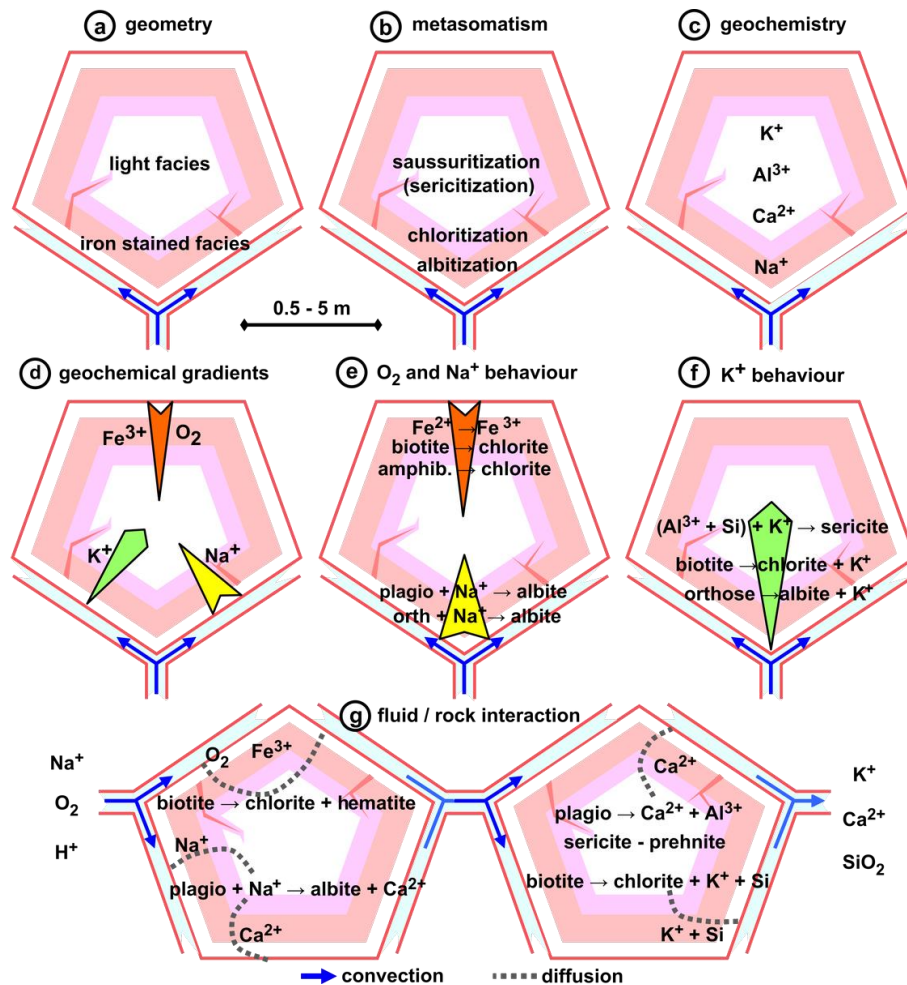
852 In this context, the coupled-reactions described above are not part of a global geochemical equilibrium
853 related to the fluids in the fractures but they are rather part of a spatially-linked dynamic geochemical

854 equilibrium. The balances are not determined by the kinetics of the mineralogical reactions but rather by the
855 kinetics of the diffusion of ions in the rock. The system was only partially open. A geochemical zoning
856 developed and progressed in a chromatographic way: with displacement of ions at diverse rates and according to
857 fronts governed by distinct reactions. This is characteristic of a metasomatic system (Korzhinskii and Oestreich
858 1965; Brinhal and Dietrich 1987; Guy 1993). Thus, thermodynamic reactions that originally seem incompatible
859 with a global context may occur. Temperature is not a limiting factor but it allows speeding up of the reactions
860 by accelerating diffusion. The reaction rate is controlled by the diffusion rate and is extremely slow, but not
861 impossible from a thermodynamical point of view.

862 Thus, within a block of massive rock bounded by fractures, a zoning of the ion transport that determines
863 the geochemical zoning is primary established (Fig. 9g). Along the fracture walls, exchanges by convection
864 prevail and are relatively fast; the mineralogical parageneses formed in equilibrium with the fluid and thereby
865 reveal the geochemical characteristics of the fluid. Further away from the fracture walls, exchanges are slow and
866 are driven by diffusion; the mineralogical parageneses are controlled by local balances that are far from the
867 balance within the fluid. This alteration of massive blocks differs entirely from alteration of porous rock where
868 ion mobility is practically exclusively provided by fluid flowthrough.

869 **7.4 Geochemistry of the fracture fluids.**

870 Albitization of plagioclase and K-feldspar reflects an interaction with Na-enriched and relatively K-
871 depleted fluids (Saigal et al. 1988; Aagaard et al. 1990; Ben Baccar et al. 1993; Schmitt 1994; Perez and Boles
872 2005). The association with hematite indicates that the fluids were also enriched in oxygen. There seems to have
873 been a relatively low CO₂ fugacity since only minor calcite was formed, in spite of the abundance of Ca released
874 during plagioclase alteration. The fluids were also relatively under-saturated in SiO₂, Al and Mg, since chlorite
875 coating the fractures is sometimes entirely altered to hematite. On exiting the system, fluids were depleted in Na
876 and O₂. However, they were enriched in K because, in the early stages of the alteration, little or no sericite was
877 formed within plagioclases during the chloritization of biotite. The chemical balance is more difficult to establish
878 for SiO₂, Al and Ca. Significant amounts of silica were liberated during chloritization of biotite so it is likely that
879 it was exported from the system, at least in the granodioritic facies that is rich in biotite. It could be that Al and
880 Ca were in balance between the altered plagioclases and the formation of prehnite and calcite, even if a
881 quantified balance is not available. Nevertheless, it is obvious that the Ca was easily exported from the cortical
882 reactional zones during the early stage of the alteration. The mobility of Al is unlikely to have occurred on a
883 significant scale because the major fractures never show any alumino-silicate mineralization. In conclusion, it
884 appears clear that the fluids responsible for the alteration were oxidized, Na-rich and relatively depleted in K in
885 comparison to Na.



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887 **Fig. 9** Schematic interpretative sketch of the various petrographic parageneses related to reddening and
 888 albitization of the granitoid rocks in the Sudetes at the outcrop scale.

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7.5 Iron oxide behavior

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Maghemite is considered to be metastable with respect to hematite (Waychunas 1991; Majzlan et al. 2003). In addition, it is generally admitted that slow and incomplete oxidation of magnetite favours the topotactic formation of maghemite whereas faster oxidation of magnetite, under highly oxidizing conditions, results in hematite formation (Steinthorsson and Helgason 1992; Bowles et al. 2011). The maghemite-hematite transition during albitization may be attributed to changes in the alteration environment during the oxidation of magnetite. In this case the alteration commenced with the formation of maghemite under low oxidizing conditions in the non-reddened and slightly reddened granitoids and continued with the formation of hematite in the reddened facies under stronger oxidizing conditions, when ferrous silicates (biotite and amphibole) were being oxidized. Thus, the occurrence of maghemite and hematite suggests that the oxidizing conditions changed during the alteration of the granitoid and the spatial distribution of hematite and maghemite provides a tool to estimate the gradient of these conditions. The distribution of the alteration environments substantiates the idea that reddened albitized facies and non reddened saussuritized/sericitized facies reflect gradational changes during a single event.

904 From a geochemical point of view there is no contradiction in the spatial discrepancy between the zones
 905 of alteration of the silicate minerals and the distribution of the iron oxides. The physico-chemical properties that
 906 constrain their reactions are not the same. A key-factor in the alteration and crystallization of the silicates is the
 907 pH, whereas Eh and the oxygen fugacity controls the oxidation state of the iron oxides. These two parameters
 908 may vary independently. Nevertheless, the precipitation of iron oxides is also facilitated by the elevated pH.
 909 Thus, local lowering of the H⁺ activity at the sites of feldspar alteration will favour the localized precipitation of
 910 hematite as inclusions within the feldspars (Boone 1969). This also explains the local distribution of pigmentary
 911 hematite inclusions in albite during the alteration of both plagioclase and K-feldspar. On the other hand, the
 912 intensity of the reddening, or the abundance of the pigmented hematite, is related to oxygen fugacity and
 913 therefore is likely to decrease away from the fractures. From the geochemical point of view, the hematite
 914 distribution is consistent with the alteration sequence of the silicate minerals.

915 **7.6 Origin of the oxidizing fluids**

916 In general, post-magmatic alterations are explained by a hydrothermal model which uses large
 917 convective circuits to bring the superficial fluids in contact with the warm rock complexes, in particular granitic
 918 intrusions, in which the alterations then take place (Taylor 1977; Jenkin et al. 1992; Putnis et al. 2007; Drake et
 919 al. 2008). The magmatic intrusions provide heat to the fluids and the subsequent loss in fluid density causes the
 920 upwelling, maintaining the circulation between the landsurface and depth. Fluids with a deeper origin may be
 921 also mixed into this circulation model.

922 In the hydrothermal model, the superficial origin of the fluids, or at least the participation of meteoric
 923 water, has been evoked by oxygen and hydrogen isotope studies on the alteration minerals (Simon and Hoefs
 924 1987; Sun and Eadington 1987; Jenkin et al. 1992; Boyce et al. 2003). These models usually consider superficial
 925 oxygen-rich fluids descending very deeply (more than 10 km) before they interact with the rock matrix.
 926 However, some reactive transport models strongly suggest that this form of fluid percolation leads to a rapid
 927 depletion of the oxygen fugacity in the circulating fluids, so fluids at great depths are very likely to be under-
 928 saturated in oxygen (MacQuarrie et al. 2010).

929 Here we have documented the **Permian-Triassic-Jurassic** age of the alteration and provided arguments
 930 that it is necessarily related to the **Post-Variscan** paleosurface and *de facto* involved infiltration of weathering
 931 solutions and/or groundwater flow.

932 **7.7 Temperature and depth of formation**

933 The $\delta^{18}\text{O}$ results **allow** to estimate the fluid temperature during plagioclase albitization and K-feldspar
 934 microclinization. Assuming a local isotopic equilibrium between minerals and the fluid at the reaction interface,
 935 the isotopic mass balance was calculated for each reaction using the equation of Taylor (1977), by (Eq. 1):

$$\delta^{18}\text{O}_f^w = \delta^{18}\text{O}_i^w + \frac{R}{W} \times (\delta^{18}\text{O}_i^m - \delta^{18}\text{O}_f^m) \quad (1)$$

936 Where $\delta^{18}\text{O}_i^w$ and $\delta^{18}\text{O}_f^w$ are the initial and final oxygen isotope compositions of the fluid,
 937 respectively, R and W the percentage of oxygen atoms in the rock (or mineral) and water, respectively, and
 938 $\delta^{18}\text{O}_i^m$ and $\delta^{18}\text{O}_f^m$ the oxygen isotope composition of the initial and final minerals, respectively. For the rock-to-
 939 water oxygen mass ratio (R/W), conservative estimates of 0.5 and 1 were applied. The applied values for the

940 oxygen isotope composition of the initial fluid ($\delta^{18}\text{O}_i$) were -8 and -12‰ (VSMOW), which are typical $\delta^{18}\text{O}$
941 values observed in deep groundwaters hosted in stable granitic basements (Stober & Bucher 1999a and b; Barth
942 2000; Kloppmann et al. 2002).

943 Two options of reactions paths were considered, (1) sequential reactions, with a fluid moving across a
944 steeped sequence of albitization and microclinization, so that the mass balance of albitization is calculated first
945 and subsequently the mass balance of microclinization is calculated using as initial fluid isotopic composition
946 the $\delta^{18}\text{O}$ value of the final fluid obtained from the mass balance of previous albitization or, (2) concomitant
947 reactions, were the fluid is renewed at the beginning of each reaction. After the isotopic mass balance estimation
948 for each reaction, a temperature was calculated for each reaction using the plagioclase-water (Eq. 2) and K-
949 feldspar-water (Eq. 3) equilibrium fractionation equations of Zheng (1993).

$$1000 \ln \alpha_{Ab-H_2O} = \frac{4.33}{T^2} \times 10^6 - \frac{6.15}{T} \times 10^3 + 1.98 \quad (2)$$

$$1000 \ln \alpha_{Kfs-H_2O} = \frac{4.32}{T^2} \times 10^6 - \frac{6.27}{T} \times 10^3 + 2.00 \quad (3)$$

950 Using the two reaction paths and the crossed combinations of the different applied values for R/W and
951 $\delta^{18}\text{O}_i$, the estimated interaction temperatures of water during albitization and microclinization range between 32
952 and 89°C (average of 60°C). Significantly, and despite the large relative variations in the values applied to R/W
953 and $\delta^{18}\text{O}_i$, and the two different reaction paths, the temperature estimates present a narrow range of variability,
954 supporting that the mineral reactions developed in by interaction with “low” temperature groundwaters. Thus,
955 considering the **Permian-Triassic** age and the relatively low temperature of the alteration, the hypothesis that the
956 alteration fluids were closely related to the paleosurface and *de facto* “vertical” infiltrations of a weathering
957 solution is a plausible possibility. Similar reaction temperatures (45 to 65°C) were obtained for plagioclase
958 albitization and K-feldspar microclinization in Permian-Triassic age red-stained albitized profiles in NE Spain
959 (Fàbrega et al. 2019), that are similar in field layout, petrographic characteristics and age to the Polish sections of
960 this study.

961 Regardless of the lack of accuracy of these temperature estimates, these estimated values show that the
962 temperatures of the reactions associated with the development of red-stained albitized rocks are significantly
963 below temperatures suggested for albitization formed under hydrothermal alteration (Cathelineau 1986; Boulvais
964 et al. 2007) and tardi-magmatic alteration (Lee and Parsons 1997; Fiebig and Hoefs 2002). Instead, the low
965 temperature range accords with the geological settings, age and association of the red-stained albitized profile
966 developed on the **Post-Variscan** paleosurface **during the Permian-Jurassic period**. We suggest that the albitized
967 profile developed a few hundred metres beneath the **Post-Variscan** paleosurface by interaction with relatively
968 ‘low’ temperature groundwaters, and was subsequently exposed at the landsurface after erosion of the uppermost
969 weathering mantle.

970 **8 PALEOENVIRONMENTAL SETTING OF THE ALBITIZATION**

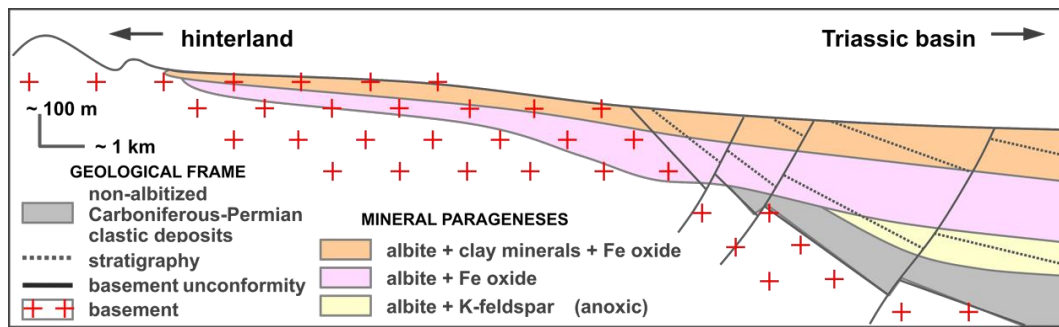
971 The absence of mesoscale- and microscale-geopetal structures, together with clues of directional water
972 flow combined with the isotropic structures around fracture blocks within the macro-scale profiles, evoke an
973 aquifer environment at the origin of the albitization. Boreholes in the Massif Central (Schmitt 1986) and field

974 exposures in the Morvan (Parcerisa et al. 2009) and NE Spain (Fàbrega et al. 2019) show that the albitized
975 profiles may generally reach 100 m in thickness and may even exceed 200 m. Such profiles are comparable to
976 those of weathering profiles with thick basal saprolite horizons in tropical soils (Boulangé 1984; Vasconcelos et
977 al. 1994; Ruffet et al. 1996; Grimaud et al. 2015) and deep groundwater saprolites in dry saline environments as
978 well as in periglacial environments (Thiry et al. 2006; Thiry et al. 2017). Nevertheless, the primary profiles were
979 probably much thicker because only the deep indurated horizons have survived the erosion that preceded the
980 **Mesozoic** transgression. The profiles were certainly at some stage surmounted by non-indurated horizons and
981 regolith materials. On the other hand, the functioning of the albitization profiles and thus landscape stability
982 spreading over ~50 Ma in the Sudetes is comparable to the duration of the deep tropical weathering profiles and
983 related landscapes rooted in old cratons mentioned above.

984 Reaction temperatures calculated from the oxygen isotope composition of minerals, and hypotheses
985 about the oxygen isotope composition of reactant fluids and their circulation, led us to an estimation of the mean
986 temperature of albitization and associated reactions of about 60°C in the Sudetes albitized profiles and 55°C in
987 the Catalan albitized profiles (Fàbrega et al. 2019). Based on this temperature, and considering a geothermal
988 gradient of about 25–30°C km⁻¹ and a surface temperature of ~30°C, in consideration of the high
989 paleotemperature proposed at the Permian–Triassic boundary, led to calculate a mean depth of about 1 km for
990 the environment in which albitization occurred.

991 There is indeed a systematic vertical profile organization for the sections we have studied, specifically
992 with a more albitized and ferruginized upper part, a less-altered and reddened lower part where albitization
993 occurred under weakly oxidizing conditions, and the progressive weakening and disappearance of alteration with
994 depth. In connecting this geometry to the **Pre-Triassic** paleosurface, it is clear that the albitization represents a
995 climatic weathering process of great magnitude. The low mobility of most of the involved chemical elements
996 could point to deep groundwater table with very low outflow. But even if the alteration occurred at a greater
997 depth, reactant solutions would have to have been renewed to provide at least sufficient O₂ and Na⁺. Therefore,
998 there must have been water flow within fractures between the fractured blocks which implies a hydraulic
999 gradient. In particular, a regional hydraulic gradient with meteoric water infiltrating the upstream profiles and
1000 groundwater outflowing downstream would have been necessary. As albitization developed, water changed
1001 composition from entrance to outflow with loss of O₂, Na⁺, and H⁺ and a likely increase in K⁺, Ca⁺⁺, and
1002 possibly SiO₂ content, as deduced from the geochemistry of the mineral transformations (Fig. 9g). The lateral
1003 evolution of such albitized profiles has been documented in the SW of the French Massif Central (Yerle and
1004 Thiry 1979; Clément 1986; Schmitt 1986). Here, (1) profiles on basement heights are thin and albitization is
1005 mainly restricted to fracture walls (such as described herein), and (2) lowland profiles rooted in Carboniferous
1006 and Permian deposits are much thicker with upper horizons albitised and oxidized whereas deeper albitized
1007 horizons remain anoxic and are the site of K-feldspar neoformations (Fig. 10).

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1010 **Fig. 10** Arrangement of mineral parageneses on the southern edge of the Massif Central. Albitized profiles
 1011 thicken from crystalline heights towards Carboniferous and Permian basin lowlands where K-feldspar forms at
 1012 depth. Modified after Clément (1986).

1013 Two fundamental elements must be taken into account: (1) the presence of relief to ensure the
 1014 hydrological gradient, and (2) the origin of sodium that triggers the whole chain of alteration. In terms of the
 1015 groundwater dynamics, it could be that albitization developed in landscapes of intermediate relief, perhaps on
 1016 large piedmonts between a mountain range and sea level. Petrographic data indicate that sodium was imported
 1017 without potassium enrichment. Early modeling attempts (Schmitt 1994) also indicated that a high Na^+/K^+ ratio
 1018 was probably required. The Na^+ enrichment is most likely to be linked with the peculiar geochemical setting of
 1019 the Permian-Triassic environments in the region. Sodium could have originated from the gigantic salt deposits of
 1020 these periods and may have been imported in large quantities and over long distances by successive phases of
 1021 eolian reworking from littoral and endoreric basins during dry periods (Bourquin et al. 2011).

1022 It should be stressed that the albitization started long before the recorded dating, possibly as early as
 1023 Mid-Permian, with the development of the Zechstein evaporite deposits in the Germanic basin. It is the final
 1024 stage that we have dated: previous stages have likely been overprinted by succeeding reactions. Classically,
 1025 weathering profiles progress from top to base and result in older ages in the uppermost horizons and younger
 1026 ages at depth, as evidenced in numerous long-lasting paleoweathering sequences (Vasconcelos et al. 1994;
 1027 Dammer 1995; Ruffet et al. 1996). However, paleomagnetic dating of the profiles we have studied here show an
 1028 inverse polarity, with the oldest ages in the weakly albitized and reddened bottom horizons and the youngest
 1029 ages in the more albitized and ferruginized uppermost horizons. This points to a 'regressive' weathering
 1030 sequence, in which the weathering profile did not continue to deepen but was 'blocked' with only the upper
 1031 horizons continuing to be weathered. This profile form relates in some way to the specific evolution of the
 1032 landscape in the region in relation to the widespread Mesozoic transgression. It would have corresponded to a
 1033 rise in basal hydrological level which induced rising watertables across the continent. The rising watertables
 1034 would have limited the infiltration of surficial waters, changed the hydrodynamic gradients and effectively
 1035 stopped the alteration at depth. Thus, the regional geodynamics during the Late Permian and Triassic period
 1036 strongly influenced weathering and regolith processes while the climate was steadily warm and semi-arid.

1037 Weathering toposequences should be considered on a continental scale as shown by Fàbrega et al.
 1038 (2019). In the case of our study area, the toposequence is arranged between the relief of the Variscan Belt that
 1039 provided the hydraulic gradient and the Permian and Triassic basins that probably supplied the salt. Albitization
 1040 was initiated and probably gradually invaded the entire **Post-Variscan** paleosurface across several 10's of Ma
 1041 during which both landscape and climate were relatively stable. Indeed, the overall ages determined up to now

1042 for the albitization span a relatively long period, between 270 and 170 Ma, that corresponds to the Middle
1043 Permian to the Middle Jurassic (Table 4). The distribution of the now-dated albitized profiles along the Post-
1044 Variscan paleosurface between the Zechstein Sea in the North and the Tethys in the South (Fig. 11), shows that:

- 1045 4) the profiles closest to the marine environments, namely the Sudetes and NE Spain, show the oldest ages
1046 (this study; Fàbrega et al. 2019);
- 1047 5) the Vosges massif, which was connected to the marine environment through the threshold passage of the
1048 Triassic transgression coming from the North, also show relatively old albitized profiles (Vercruysse 2011);
- 1049 6) the area most distant from the sea, namely the Rouergue heights overlay by infra-Jurassic deposits,
1050 comprises the youngest albitized profiles (Schmitt et al. 1984) while the Permian graben covered with
1051 coarse Triassic deposits on lower plateaux in Rouergue (Bonhomme et al. 1980) and Morvan (Ricordel et al.
1052 2007) have older ages; and
- 1053 7) in the Sudetes, the highly evolved ferruginous albitized facies currently in hillside positions are relatively
1054 young while the valley-bottom sections are older. This differentiation is possibly related to local
1055 morphologies whereby raised tectonic blocks remained emerged during the Mesozoic transgression.

1056 The concordance between age and relative position in the Triassic paleogeography is striking. The
1057 areas covered early by the Triassic transgression have the oldest albitized profiles whereas the highlands distant
1058 from the sea have the youngest ages. It is actually the interruption of weathering by the Triassic transgression
1059 that is dated. These datings could in future be used as a tool to better enhance the mapping of the Triassic
1060 transgression and the geodynamics of the basement as well as its post-Triassic evolution.

1061 Paleolandscape and paleoweathering features of the Post-Variscan paleosurface in Western Europe may
1062 be compared to regolith features existing today on the African, Australian and South American cratons inherited
1063 from Pangea break out. Combined radiogenic dating and geomorphological analyses allowed us to document
1064 paleolandscape and paleoweathering profiles that functioned for more than 100 Ma in these cratonic regions
1065 (Vasconcelos et al. 1994; Dammer 1995; Ruffet et al. 1996; Pillans, 2007; Beauvais et al. 2008; Peulvast et al.
1066 2008 ; Beauvais and Chardon 2013; Jean et al. 2020). This is the spatial and temporal scale of the albitization
1067 and hematization paleoweathering profiles that we studied on the Western European Post-Variscan paleosurface.
1068 Persistence of the Post-Variscan paleosurface is conformed by the widespread magnetic overprinting phase due to
1069 secondary hematite formation (related or not to albitization) that affected widely the Variscan belt during the
1070 235-250 Ma time-span Edel et al. (2018).

1071

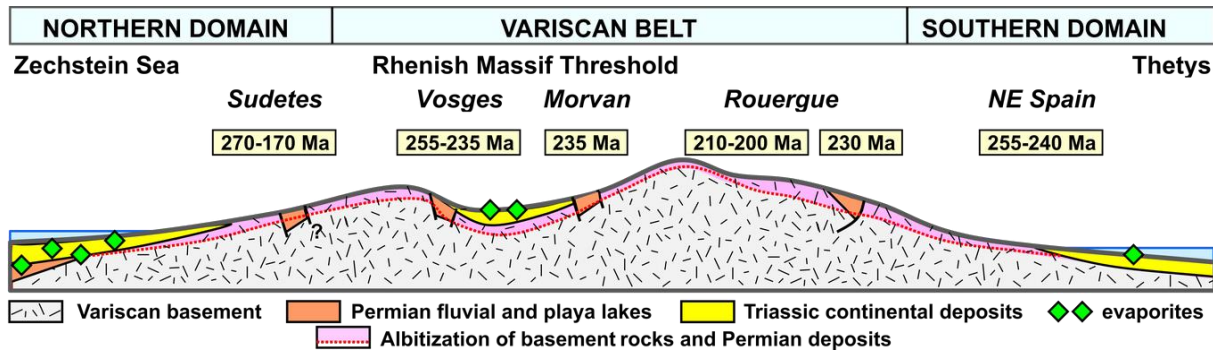
1072 Table 4 – Age of the reddened/albitized rocks related to the Post-Variscan paleosurface on the Variscan Belt.

1073

area	location	geomorph	rockl	sample	method	age	reference
Sudetes	Szklarska Poręba	hill	granitoide	2.5 x 2.5 cm	palaeomag	256-190	herein
	Laski quarry	valley	granitoide	2.5 x 2.5 cm	palaeomag	270-190	herein
	Laski village	hill slope	granitoide	2.5 x 2.5 cm	palaeomag	258-170	herein
	Chwalisław	valley	granitoide	2.5 x 2.5 cm	palaeomag	205-195	herein
	Chwalisław	valley	granitoide	monazite	U-Th-Pb _{tot}	254	herein
	Chwalisław	valley	granitoide	2.5 x 2.5 cm	palaeomag	250	Edel et al. (1997)
Vosges			granitoide	2.5 x 2.5 cm	palaeomag	255-235	Vercruysse (2011)
Morvan			granitoide	2.5 x 2.5 cm	palaeomag	235	Ricordel et al. (2007)
Rouergue	Bertholène	high basin	orthogneiss sediments	minerals	K-Ar	210-200	Schmitt et al. (1984)
	Brousse			clay fraction	K-Ar	230	Bonhomme et al. (1980)
NE Spain	Guilleries Massif		granite	monazite	U-Th-Pb _{tot}	251	Fàbrega et al. (2019)
	Roc Frausa Mass.		granite	monazite	U-Th-Pb _{tot}	254	Fàbrega et al. (2019)
	Hortmoier Massif		granite	K-F crystals	K-Ar	240	Fàbrega et al. (2019)

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1078 **Fig. 11** Position and age of the studied albitized profiles on a schematic N-S geomorphological section of the
1079 **Post-Variscan paleosurface** and associated sedimentary basins during Permian-Triassic times. The oldest ages
1080 occur nearest the sea and youngest away the sea and in highland areas that were later buried. Paleogeography
1081 after Bourquin et al. (2011).

1082 9 CONCLUSIONS

1083 The oxidized and albitized facies of the Polish Sudetes granitoids exhibit two types of ‘classical’
1084 petrographical parageneses: a saussuritization-sericitization and an albitization-hematitization. The chloritization
1085 of amphiboles and biotites that is associated to these alterations is intermediate between both. These alterations
1086 are always closely associated without any break or spatial discordance between them. They are controlled by the
1087 presence of fractures and are spatially arranged with respect to the configuration of these fractures. Albitization
1088 and hematitization are preferentially intense close to the walls of the fractures whereas the saussuritization and
1089 the sericitization of plagioclase become more important further away from the fractures and towards the center
1090 of the massive blocks. These alterations can be considered as topochemical. However, the complete albitization
1091 of the plagioclase and K-feldspar in the vicinity of the fractures signifies a supply of Na. Hematite associated
1092 with albite is an indication of oxidative capacity and therefore the supply of oxygen via the alteration.

1093 Paleomagnetic dating of the reddened and albitized facies reveal **Late Permian to Early Jurassic** ages for
1094 the iron-oxides and radiochronological dating of associated monazite using the U-Th-Pb_{total} method indicates
1095 **Late Permian** recrystallization ages. Given the ages of the alteration, it can clearly be related to the extensive
1096 **Post-Variscan** paleosurface in the region, since the altered granitoids were already outcropping during from **Late**
1097 **Permian to Early Jurassic**. Therefore, these reddened and albitized profiles can be considered a consequence of
1098 deep weathering.

1099 As there was clearly a requirement to renew solutions to feed chemical reactions, it is likely that the
1100 alteration profiles were downstream of highlands in order to generate a hydrological head. The sodium supply
1101 that triggered albitization could well have come from the gigantic salt playas that characterized the lowland
1102 paleogeography of the Permian and Triassic in the region. Thus, we suggest that the reddened and albitized
1103 profiles developed on a wide glacia shaped in the basement rocks upstream of the evaporatic basins. The deep
1104 weathering was interrupted and preserved by the **Mesozoic** transgression.

1105 The age dating shows that the oldest examples of the alteration profiles (Late Permian to Early Triassic)
1106 are proximal to the Zechstein and Tethys seas, while the youngest (late Triassic to basal Jurassic) are further

1107 away and appear to be located on the highlands that were more recently covered by the Mesozoic transgression.
1108 Thus, this alteration of the crystalline rocks is an important marker to constrain Triassic and post-Triassic
1109 geodynamics in the region, including erosion rates and the geomorphic evolution of the Paleozoic massifs.

1110 Albitized and oxidized facies of crystalline basement are widespread throughout the world. Until now,
1111 this subject has attracted relatively little interest from petrographers. A re-examination of all existing data could
1112 involve a reconsideration of questions such as prehnite and sericite paragenesis, particularly from the viewpoint
1113 of temperature of formation, since they have in the past been used as an indicator of ‘high temperatures’. It may
1114 be that the formation of sericite and prehnite is not part of a unique alteration leading to albitization, but the fact
1115 remains that proximity to the paleosurface is clearly indicated by the presence of hematite.

1116 We suggest that particular attention should be paid to the radiochronological and paleomagnetic dating
1117 results that some have considered to be “obviously too young” in the context of the age of emplacement of the
1118 crystalline massifs. These “too young” dates have been systematically rejected as false in the past and rarely
1119 noted in publications; consequently this information was lost. A new consideration of these dates may realise
1120 their consistency, as shown for example by Edel and Schneider (1995) for the Permian formations that endured a
1121 Triassic rejuvenation highlighted by paleomagnetism.

1122 It should be noted that the formation of secondary apatite in the reddened albitized Sudetes rocks has
1123 previously been reported in conjunction with albitization and chloritization related to the Triassic paleosurface in
1124 the French Massif Central (Parcerisa et al. 2009) and in NE Spain (Fàbrega et al. 2019). It also formed during
1125 metasomatic albitization in a metagabbro from Norway (Engvik et al. 2009). The formation of significant
1126 amounts of secondary apatite during albitization must be taken into account if fission track analysis or (U–
1127 Th)/He dating is carried out on these reddened albitized rocks.

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1129 **References:**

1130

1131 Aagaard P, Egeberg PK, Saigal GC, Morad S, Bjørlykke K (1990) Diagenetic albitization of detrital K-feldspars
1132 in Jurassic, Lower Cretaceous, and Tertiary clastic reservoir rocks from offshore Norway, II. Formation
1133 water chemistry and kinetic considerations. *Journal of Sedimentary Petrology* 60:575–581.

1134 Aleksandrowski P, Kryza R, Mazur S, Żaba J (1997) Kinematic data on major Variscan strike-slip faults and
1135 shear zones in the Polish Sudetes, northeast Bohemian massif. *Geological Magazine* 133: 727-739.

1136 Aleksandrowski P, Mazur S (2002) Collage tectonics in the northeasternmost part of the Variscan Belt: the
1137 Sudetes, Bohemian Massif). In: Winchester, J., Pharaoh T. & Verniers J. (eds), *Palaeozoic Amalgamation*
1138 of Central Europe, Geological Society, London, Special Publications 201: 237-277.

1139 Alt, J.C., Laverne, C., Coggon, R.M., Teagle, D.A., Banerjee, N.R., Morgan, S., Smith-Duque, C.E., Harris, M.,
1140 Galli, L. (2010) Subsurface structure of a submarine hydrothermal system in ocean crust formed at the
1141 East Pacific Rise, ODP/IODP Site 1256. *Geochemistry, Geophysics, Geosystems*, 11(10).

1142 Awdankiewicz M, Awdankiewicz H (2010) Spessartites of the Kłodzko Złoty-Stok dyke swarm. *Mineralogia*
1143 Special Papers 37: 135-139.

- 1144 Bachliński R, Bagiński B (2007) Kłodzko-Złoty-Stok granitoids massif. In Kozłowski A., Wiszniewska J (eds)
1145 In Granitoids in Poland, *Archiwum mineralogiczne*, Monograph 1: 261-273.
- 1146 Baranowski Z, Haydukiewicz A, Kryza R, Lorenc S, Muszynski A (1990) Outline of the geology of the Góry
1147 Kaczawskie (Sudetes, Poland). *Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen* 179: 223-
1148 257.
- 1149 Barth SR (2000) Geochemical and boron, oxygen and hydrogen isotopic constraints on the origin of salinity in
1150 groundwaters from the crystalline basement of the Alpine Foreland. *Applied Geochemistry* 15:937–952.
1151 [https://doi.org/10.1016/S0883-2927\(99\)00101-8](https://doi.org/10.1016/S0883-2927(99)00101-8)
- 1152 Beauvais A, Chardon F (2013) Modes, tempo, and spatial variability of Cenozoic cratonic denudation: The West
1153 African example, *Geochem. Geophys. Geosyst.*, 14, 1590–1608, doi:10.1002/ggge.20093.
- 1154 Beauvais, A., Ruffet G, Henocque O, Colin F (2008) Chemical and physical erosion rhythms of the West
1155 African Cenozoic morphogenesis: The ³⁹Ar-⁴⁰Ar dating of supergene K-Mn oxides, *J. Geophys. Res.*,
1156 113, F04007, doi:10.1029/2008JF000996.
- 1157 Ben Baccar M, Fritz B, Madé B (1993) Diagenetic albitization of K-feldspar and plagioclase in sandstone
1158 reservoirs: thermodynamic and kinetic modeling. *Journal of Sedimentary Petrology* 63:1100–1109.
- 1159 Besse J, Courtillot V (2003) Apparent true polar wander and the geometry of the geomagnetic field over the last
1160 200 Myr: Correction. *Journal of Geophysical Research* 108: 2300.
- 1161 Bonhomme M, Yerle J.-J, Thiry M (1980) Datation K-Ar de fractions fines associées aux minéralisations. Le cas
1162 du bassin uranifère permio-houiller de Brousse-Broquiès (Aveyron, France).- C.R. Académie Sciences
1163 Paris 291D: 121-123.
- 1164 Boone GM (1969) Origin of clouded red feldspars, petrology contrasts in a granitic porphyry intrusion.
1165 *American Journal of Science* 267: 633-668.
- 1166 Borkowska M, (1966) Petrografia granitu Karkonoszy (Petrography of the Karkonosze Granite). *Geologia*
1167 *Sudetica* 2: 7-119. (in Polish with English extended summary)
- 1168 Botor D, Dunkl I, Anczkiewicz A, Mazur S (2017) Post-Variscan thermal history of the Moravo-Silesian lower
1169 Carboniferous Culm Basin (NE Czech Republic - SW Poland). *Tectonophysics* 712-713: 643-662.
- 1170 Boulangé B (1984) Les formations bauxitiques latéritiques de Côte d'Ivoire. *Travaux et Documents ORSTOM*
1171 175, 341 p.
- 1172 Boulvais P, Ruffet G, Cornichet J, Mermet M (2007) Cretaceous albitization and dequartzification of Hercynian
1173 peraluminous granite in the Salvezines Massif (French Pyrénées). *Lithos* 93: 89-106.
- 1174 Bourquin S, Bercovici A, López-Gómez J, Diez J-B, Broutin J, Ronchi A, Amour F (2011) The Permian–Triassic
1175 transition and the onset of Mesozoic sedimentation at the northwestern peri-Tethyan domain scale:
1176 palaeogeographic maps and geodynamic implications. *Palaeogeography Palaeoclimatology*
1177 *Palaeoecology* 299: 265-280.
- 1178 Bowles JFW, Howie RA, Vaughan DJ, Zussman J (2011) Non-Silicates: Oxides, Hydroxides and Sulphides.
1179 *Rock-Forming Minerals*, 2nd edn, Geological Society of London 5A: 920 p.

- 1180 Boyce AJ, Fulignati P, Sbrana A (2003) Deep hydrothermal circulation in a granite intrusion beneath Larderello
1181 geothermal area (Italy): constraints from mineralogy, fluid inclusions and stable isotopes. *Journal of*
1182 *Volcanology and Geothermal Research* 126: 243-262.
- 1183 Brinhall G, Dietrich W (1987) Constitutive mass balance relations between chemical composition, volume,
1184 density, porosity, and strain in metasomatic hydrochemical systems: Results on weathering and
1185 pedogenesis. *Geochimica et Cosmochimica Acta* 51: 567-587.
- 1186 Butler RF (1998) *Paleomagnetism: Magnetic domains to geologic terranes*. Electronic edition, 237 p.
- 1187 Castet S, Dandurand J-L, Schott J, Gout R (1993) Boehmite solubility and aqueous aluminium speciation in
1188 hydrothermal solutions (90-350°C): Experimental study and modeling. *Geochimica et Cosmochimica*
1189 *Acta* 57: 4869-4884.
- 1190 Cathelineau M (1986) The hydrothermal alkali metasomatism effects on granitic rocks: quartz dissolution and
1191 related subsolidus changes. *Journal of Petrology* 27: 945-965.
- 1192 Chrzęstek A., Wojewoda J (2011) Mezozoik południowo-zachodniej Polski - synklinorium północnosudeckie
1193 (Mesozoic of SW Poland, the North Sudetic Synclinorium). In: Żelaźniewicz, A., Wojewoda, J.,
1194 Ciężkowski, W. (eds), *Mezozoik i kenozoik Dolnego Śląska*. Polskie Towarzystwo Geologiczne,
1195 Wrocław, 1-10 (in Polish, English abstract).
- 1196 Clément J-Y, (1986) *Minéralogie, pétrologie et géochimie du Permien de Lodève (Hérault, France) : diagenèse*
1197 *précoce, altération feldspathisante et mise en place des minéralisations uranifères*. E.N.S.M.P. Mémoire
1198 *Sciences de la Terre* 2: 137 p.
- 1199 Cocherie A, Albarede F (2001) An improved U-Th-Pb age calculation for electron microprobe dating of 845
1200 monazite. *Geochimica et Cosmochimica Acta* 65:4509–4522, [https://doi.org/10.1016/S0016-](https://doi.org/10.1016/S0016-7037(01)00753-0)
1201 [7037\(01\)00753-0](https://doi.org/10.1016/S0016-7037(01)00753-0) 846
- 1202 Cocherie A, Legendre O, Peucat J-J, Kouamelan AN (1998) Geochronology of polygenetic monazites 847
1203 constrained by in situ electron microprobe Th-U-total lead determination: Implications for lead 848
1204 behaviour in monazite. *Geochimica et Cosmochimica Acta* 62: 2475-2497,
1205 [https://doi.org/10.1016/S0016-7037\(98\)00171-9](https://doi.org/10.1016/S0016-7037(98)00171-9)
- 1206 Cogné J-P (2003) PaleoMac: a Macintosh™ application for treating paleomagnetic data and making plate
1207 reconstructions. *Geochemistry Geophysics Geosystems* 4: 1007.
- 1208 Dammer D (1995) *Geochronology of chemical weathering processes in the northern and western Australian*
1209 *regolith*. PhD thesis, Australian National University: 214 p.
- 1210 Danišik M, Migoń P, Kuhlemann J, Evans NJ, Dunkl I, Frisch W (2010) Thermochronological constraints on the
1211 long-term erosional history of the Karkonosze Mts, central Europe, *Geomorphology* 117: 78–89.
- 1212 Danišik M, Štěpančíková P, Evans NJ (2012) Constraining long-term denudation and faulting history in
1213 intraplate regions by multisystem thermochronology: An example of the Sudetic Marginal Fault
1214 (Bohemian Massif, central Europe). *Tectonics* 31: TC2003, <https://doi.org/10.1029/2011TC003012>

- 1215 De Jong G., Williams PJ (1995). Giant metasomatic systems formed during exhumation of mid-crustal
1216 Proterozoic rocks in the vicinity of the Cloncurry fault, northwest Queensland. *Australian Journal of*
1217 *Earth science* 42: 281-290.
- 1218 De La Roche H (1957) *Eléments pour l'étude de la saussuritisation. Mémoires de l'Institut Scientifique de*
1219 *Madagascar* 7D: 217-223.
- 1220 Depciuch T (1972) Wiek bezwzględny (K-Ar) granitoidów Kłodzko Złotostockich i strefy Niemczy. *Kwartalnik*
1221 *Geologiczny* 16: 103-111.
- 1222 Drake H, Tullborg EL, Annersten H (2008) Red-staining of the wall rock and its influence on the reducing
1223 capacity around water conducting fractures. *Applied Geochemistry* 23: 1898-1920.
- 1224 Dziedzic K, Teisseyre AK (1990) The Hercynian molasse and younger deposits in the Intra-Sudetic Depression,
1225 SW Poland. *Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen* 179: 285–305.
- 1226 Edel J-B, Schneider J-L (1995) The Late Carboniferous to Early Triassic geodynamics evolution of the Variscan
1227 Europe in the light of magnetic overprints in Early Permian rhyolites from the northern Vosges (France)
1228 and the central Black Forest (Germany). *Geophysical Journal International* 122: 858-876.
- 1229 Edel J-B, Düringer P (1997) The apparent polar wander path of the European plate in Upper Triassic-Lower
1230 Jurassic times and the Liassic intraplate fracturing of the Pangea: New palaeomagnetic constraints from
1231 NW France and SW Germany. *Geophysical Journal International* 128: 331-344.
- 1232 Edel J-B, Aifa T, Jelenska M, Kadzialko-Hofmohl M, Zelazniewicz A (1997) Réaimantations des formations
1233 paléozoïques des Sudètes polonaises et courbe de dérive des pôles géomagnétiques d'Europe du
1234 Carbonifère moyen au Jurassique moyen. *Comptes Rendus de l'Académie des Sciences - Séries IIA -*
1235 *Earth and Planetary Science Letters* 325: 479-486.
- 1236 Edel J B, Schulmann K, Lexa O, Lardeaux J-M. (2018) Late Palaeozoic palaeomagnetic and tectonic constraints
1237 for amalgamation of Pangea supercontinent in the European Variscan belt. *Earth-science reviews*, 177:
1238 589-612
- 1239 Engvik AK, Golla-Schinder U, Berndt J, Austrheim H, Putnis A (2009) Intragranular replacement of chlorapatite
1240 by hydroxy-fluor-apatite during metasomatism. *Lithos* 112: 236-246.
- 1241 Engvik AK, Putnis A, Gerald JDF, Austrheim H (2008) Albitisation of granitic rocks: the mechanism of
1242 replacement of oligoclase by albite. *The Canadian Mineralogist* 46: 1401-1415.
- 1243 Fàbrega C, Parcerisa D, Rossell JM, Gurenko A, Franke C (2017) Predicting instrumental mass fractionation
1244 (IMF) of stable isotope SIMS analyses by response surface methodology (RSM). *J Anal At Spectrom*
1245 32:731–748. <https://doi.org/10.1039/C6JA00397D>
- 1246 Fàbrega C, Parcerisa D, Thiry M, Franke C, Gurenko A, Gómez-Gras D, Solé J, Travé A. (2019) Permian–
1247 Triassic red-stained albitized profiles in the granitic basement of NE Spain: evidence for deep alteration
1248 related to the Triassic palaeosurface. *International Journal of Earth Sciences* 108: 2325-2347.

- 1249 Farmer VC, Lumsdon DG (1994) An assessment of complex formation between aluminium and silicic acid in
1250 acidic solutions. *Geochimica et Cosmochimica Acta* 58: 3211-3334.
- 1251 Feist-Burkhardt S, Götz A, Szulc J, Borkhataria R, Geluk M, Haas J, Hornung J, Jordan P, Kempf O, Michalík J,
1252 Nawrocki J, Reinhardt L, Ricken W, Röhling H-G, Ruffer T, Török Á, Zühlke R (2008) Triassic. In:
1253 McCann (ed.), *The Geology of Central Europe, Volume 2, Mesozoic and Cenozoic*. Geological Society
1254 of London, pp 749-821.
- 1255 Ferry JM (1979) Reaction mechanisms, physical conditions, and mass transfer during hydrothermal alteration of
1256 mica and feldspar in granitic rocks from south-central Maine, USA. *Contributions to Mineralogy and
1257 Petrology* 68:125-139.
- 1258 Fiebig J, Hoefs J (2002) Hydrothermal alteration of biotite and plagioclase as inferred from intragranular oxygen
1259 isotope-and cation-distribution patterns. *European Journal of Mineralogy*, 14, 49-60.
- 1260 Fisher RA (Ed.) (1953) *Dispersion on a Sphere*. Philosophical Transactions of the Royal Society of London.
1261 Series A, vol. 217. 295 pp.
- 1262 Fisher NI, Lewis T, Embleton BJJ (1987) *Statistical Analysis of Spherical Data*, Cambridge Univ. Press,
1263 Cambridge 1-329.
- 1264 Głuszyński, A., Aleksandrowski, P. (2022) Late Cretaceous–Early Palaeogene inversion-related tectonic
1265 structures at the northeastern margin of the Bohemian Massif (southwestern Poland and northern
1266 Czechia). *Solid Earth*, 13(8): 1219-1242.
- 1267 Goldsmith JR, Laves F (1954) The microcline-sanidine stability relations. *Geochimica et Cosmochimica Acta* 5:
1268 1-19.
- 1269 González-Acebrón L, Arribas J, Mas R (2010) Role of sandstone provenance in the diagenetic albitization of
1270 feldspars. A case study of the Jurassic Tera Group sandstone (Camos Basin, NE Spain). *Sedimentary
1271 Geology* 229: 53-63.
- 1272 González-Acebrón L, Götze J, Barca D, Arribas J, Mas R, Perez-Garrido C (2012) Diagenetic albitization in the
1273 Tera Group, Camos Basin (NE Spain) recorded by trace elements and spectral cathodoluminescence.
1274 *Chemical Geology* 312, 148-162.
- 1275 Götze J., Krbetschek M.R., Habermann D., Wolf D. 1999: High-resolution cathodoluminescence studies of
1276 feldspar minerals. [In]: *Cathodoluminescence in geosciences*. Ed.: Pagel M., Barbin V., Blanc Ph.,
1277 Ohnenstetter D., Springer-Verlag, Berlin Heidelberg New York: 271-302.
- 1278 Götze J, Krbetschek MR, Habermann D, Wolf D (2000) High-resolution cathodoluminescence studies of
1279 feldspar minerals. In: *Cathodoluminescence in geosciences*. Ed.: Pagel M., Barbin V., Blanc Ph.,
1280 Ohnenstetter D., Springer-Verlag, Berlin Heidelberg New York: 245-270.
- 1281 Grimaud J-L, Chardon, Metelka V, Beauvais A, Bamba O (2015) Neogene cratonic erosion fluxes and landform
1282 evolution processes from regional regolith mapping (Burkina Faso, West Africa). *Geomorphology* 241:
1283 315-330.

- 1284 Guy B (1993) Mathematical revision of Korzhinskii's theory of infiltration metasomatic zoning. *European*
1285 *Journal of Mineralogy* 5: 317-339.
- 1286 Hövelmann J, Putnis A, Geisler T, Schmidt BC, Golla-Schindler U (2010) The replacement of plagioclase
1287 feldspars by albite: observations from hydrothermal experiments. *Contribution to Mineralogy and*
1288 *Petrology* 159: 43-59.
- 1289 **Jean A, Beauvais A, Chardon D, Arnaud N, Jayananda M, Mathe PE (2020) Weathering history and landscape**
1290 **evolution of Western Ghats (India) from $^{40}\text{Ar}/^{39}\text{Ar}$ dating of supergene K–Mn oxides. *Journal of the***
1291 ***Geological Society*, 177(3): 523-536.**
- 1292 Jenkin GRT, Fallick AE, Leake BE (1992) A stable isotope study of retrograde alteration in SW Connemara,
1293 Ireland. *Contributions to Mineralogy and Petrology* 110: 269-288.
- 1294 Jokubauskas P, Bagiński B, MacDonald R, Awdankiewicz M (2014) Field trip to the Kłodzko – Złoty Stok
1295 intrusion. *Mineralogia, Mineralogical Society of Poland, Kraków, Special Papers* 42: 127-134.
- 1296 Jokubauskas P, Bagiński B, Macdonald R, Krzemińska E (2018) Multiphase magmatic activity in the Variscan
1297 Kłodzko–Złoty Stok intrusion, Polish Sudetes: evidence from SHRIMP U–Pb zircon ages. *International*
1298 *Journal of Earth Sciences* 107:1623–1639.
- 1299 Kadzialko-Hofmokl M, Kruczyk J, Mazur S, Siemiakowski J (2003) Paleomagnetism of the Upper Paleozoic
1300 and devonian rocks from the Kłodzko metamorphic Complex in the West Sudetes (SWPoland): tectonic
1301 implications for the Variscan Belt of Central Europe. *Tectonophysics* 177 : 83-99.
- 1302 **Kelley, D.S., Robinson, P.T., Malpas, J.G. (1992) Processes of brine generation and circulation in the oceanic**
1303 **crust: fluid inclusion evidence from the Troodos ophiolite, Cyprus. *Journal of Geophysical Research:***
1304 ***Solid Earth*, 97(B6): 9307-9322.**
- 1305 Kirschvink JL (1980) The least-square line and plane and the analysis of paleomagnetic data, *Geophys. J. R.*
1306 *Astron. Soc.* 62: 699-718.
- 1307 Kloppmann W, Girard J-P, Négrel P (2002) Exotic stable isotope compositions of saline waters and brines from
1308 the crystalline basement. *Chemical Geology* 184: 49–70. [https://doi.org/10.1016/S0009-2541\(01\)00352-7](https://doi.org/10.1016/S0009-2541(01)00352-7)
- 1309 Korzhinskii DS, Oestreich W (1965) *Abriss der metasomatischen Prozesse*. Akademie-Verlag Berlin: 195 p.
- 1310 Kowalski A (2020) Triassic palaeogeography of NE Bohemian Massif based on sedimentological record in the
1311 Wleń Graben and the Krzeszów Brachysyncline (SW Poland). *Annales Societatis Geologorum Poloniae*
1312 90: 125–148.
- 1313 Kozłowski A, Karwowski Ł, Olszyński W (1975) Tungsten-tin-molybdenum mineralization in the Karkonosze
1314 massif. *Acta Geologica Polonica* 25: 415-430.
- 1315 Kroner U, Mansy J-L, Mazur S, Aleksandrowski P, Hann HP, Huckriede H, Lacquement F, Lamarche J, Ledru
1316 P, Pharaoh T, Zedler H, Zeh A, Zulauf G (2008) Variscan Tectonics. In: McCann, T. (ed.): *The Geology*
1317 *of Central Europe*. The Geological Society, London: 599-664.

- 1318 Kryza R, Crowley QG, Larionov A, Pin Ch, Oberc-Dziedzic T, Mochnacka K (2012) Chemical abrasion applied
1319 to SHRIMP zircon geochronology: An example from the Variscan Karkonosze Granite (Sudetes, SW
1320 Poland). *Gondwana Research* 21: 757-767.
- 1321 Kryza R, Schaltegger U, Oberc- Dziedzic T, Pin Ch, Ovtcharova M (2014) Geochronology of a composite
1322 granitoid pluton: a high- precision ID- TIMS U–Pb zircon study of the Variscan Karkonosze Granite
1323 (SW Poland). *International Journal of Earth Sciences* 103: 683-696.
- 1324 Kusiak M.A, Williams IS, Dunkley DJ, Konečný P, Slaby E, Martin HM (2014) Monazite to the rescue: U-Th-
1325 Pb dating of the intrusive history of the composite Karkonosze pluton, Bohemian Massif. *Chemical*
1326 *Geology* 364: 76-92.
- 1327 Lee MR, Parsons I (1997) Dislocation formation and albitization in alkali feldspars from the Shap granite.
1328 *American Mineralogist*, 82: 557-570.
- 1329 Leichmann J, Broska I, Zachovalová K (2003) Low- grade metamorphic alteration of feldspar minerals: a CL
1330 study. *Terra Nova*, 15: 104-108.
- 1331 Lewandowski, M., Werner, T., Nowoz'yn'ski, K., 1997. PDA—a package of Fortran programs for
1332 paleomagnetic data analysis. *Inst. Geophys. Pol. Ac. Sci.* (Personal communication).
- 1333 Lorenc M (1994) Rola magm zasadowych w ewolucji intruzji granitoidowych (studium porównawcze
1334 wybranych masywów hercyńskich). *Geologia Sudetica* 28: 1-130.
- 1335 Ludwig KR (2003) User's Manual for Isoplot 3.00 - A Geochronological Toolkit for Microsoft Excel. 940
1336 Berkeley Geochronological Centre Spec Publ Special Pu: 25–32
- 1337 McFadden PL, McElhinny MW (1988) The combined analysis of remagnetization circles and direct observations
1338 in paleomagnetism, *Earth Plan. Sci. Letts.* 87: 161-172.
- 1339 Machowiak K, Armstrong R (2007) SHRIMP U-Pb zircon age from the Karkonosze granite. *Mineralogia*
1340 *Polonica Special Papers* 31: 193-196.
- 1341 MacQuarrie KTB, Mayer KU, Jin B, Spiessl SM (2010) The importance of conceptual models in the reactive
1342 transport simulation of oxygen ingress in sparsely fractured crystalline rock. *Journal of contaminant*
1343 *hydrology* 112: 64-76.
- 1344 Majzlan J, Grevel KD, Navrotsky A (2003) Thermodynamics of Fe oxides: Part II. Enthalpies of formation and
1345 relative stability of goethite (α -FeOOH), lepidocrocite (γ -FeOOH), and maghemite (γ -Fe₂O₃). *American*
1346 *Mineralogist* 88: 855-859.
- 1347 Marheine D., Kachlik V., Maluski H., Patočka F., Żelaźniewicz A. 2002. The ⁴⁰Ar/³⁹Ar ages from the West
1348 Sudetes (NE Bohemian Massif): constraints on the Variscan polyphase tectonothermal development.
1349 *Geol. Soc. Spec. Publ.*, 201, 133-155
- 1350 Marshall DJ (1988) *Cathodoluminescence of Geological Materials*. Unwin Hyman, Boston, 146 p.
- 1351 Matte Ph, Maluski H, Rajlich P, Franke W (1990) Terrane boundaries in the Bohemian Massif: results of large
1352 scale Variscan shearing. *Tectonophysics* 177: 151–170.

- 1353 Matthews A, Goldsmith JR (1984) The influence of metastability on reaction kinetics involving zoisite
1354 formation from anorthite at elevated pressures and temperatures. *American Mineralogist* 69: 848-857.
- 1355 Mazur S, Aleksandrowski P, Kryza R, Oberc-Dziedzic T (2006) The Variscan orogen in Poland. *Geological*
1356 *Quarterly* 50: 89-118.
- 1357 Mazur S, Aleksandrowski P, Turniak K, Awdankiewicz M (2007) Geology, tectonic evolution and late Paleozoic
1358 magmatism of the Sudetes-an overview. In Kozłowski A., Wiszniewska J (eds). In *Granitoids in Poland*.
1359 *Archiwum mineralogiczne Monograph* 1: 59-87.
- 1360 Migoń P, Lidmar-Bergström K (2001) Weathering mantles and their significance for geomorphological
1361 evolution of central and northern Europe since the Mesozoic. *Earth-Science Reviews* 56: 285–324.
- 1362 Migoń P, Danišik M (2012) Erosional history of the Karkonosze Granite Massif – constraints from adjacent
1363 sedimentary basins and thermochronology. *Geological Quarterly* 56: 440–454.
- 1364 Mikulski SZ, Williams IS, Baginski B (2013) Early Carboniferous (Visean) emplacement of the collisional
1365 Kłodzko–Złoty-Stok granitoids (Sudetes, SW Poland): constraints from geochemical data and zircon U–
1366 Pb ages. *International Journal of Earth Sciences* 102: 1007-1027.
- 1367 Mikulski S.Z, Bagiński B, Dzierżanowski P (2004) The CHIME age calculations on monazite and xenotime in
1368 aplogranite from the Szklarska Poręba Huta. *Miner. Ass. Pol. Spec. Pap.*, 24: 287–290.
- 1369 Milewicz J., 1968. Karbon górny: niecka północnosudecka. In: *Budowa Geologiczna Polski*, t.1, cz. 1. Prekambr
1370 i paleozoik. Wyd. Geologiczne, Warszawa, 439-441
- 1371 Milewicz J., 1970. The Cretaceous of the Jerzmanice Graben (Sudetes). *Biul. Inst. Geol.*, 239, 37-66
- 1372 Morad S, El-Ghaly MAK, Caja MA, Sirat M, Al-Ramadan K, Mansurbeg H (2010) Hydrothermal alteration of
1373 plagioclase in granitic rocks from Proterozoic basement of SE Sweden. *Geological Journal* 45: 105-116.
- 1374 Oelkers EH, Schott J, Devidal JL (1994) The effect of aluminum, pH, and chemical affinity on the rates of
1375 aluminosilicate dissolution reactions. *Geochimica et Cosmochimica Acta* 58: 2011-2024.
- 1376 Parcerisa D, Thiry M, Schmitt J-M (2009) Albitisation related to the Triassic unconformity in igneous rocks of
1377 the Morvan Massif (France). *International Journal of Earth Science* 99: 527-544.
- 1378 Perez RJ, Boles JR (2005) An empirically derived kinetic model for albitization of detrital plagioclase. *American*
1379 *Journal of Science* 305: 312-343.
- 1380 Peulvast J-P, Sales VC, Bétard F, Gunnell Y (2008) Low post-Cenomanian denudation depths across the
1381 Brazilian Northeast: implications for long-term landscape evolution at a transform continental margin.
1382 *Global and Planetary Change*, 62(1-2): 39-60.
- 1383 Pillans B (2007) Pre- Quaternary landscape inheritance in Australia. *Journal of Quaternary Science: Published*
1384 *for the Quaternary Research Association*, 22(5): 439-447.
- 1385 Plümper O., Putnis A (2009) The Complex Hydrothermal History of Granitic Rocks: Multiple Feldspar
1386 Replacement Reactions under Subsolidus Conditions. *Journal of Petrology* 50(5): 967-987.

- 1387 Powolny, T., Dumańska-Słowik, M., Anczkiewicz, A. A., & Sikorska-Jaworowska, M. (2022). Origin and
1388 timing of spilitic alterations in volcanic rocks from Głuszyca Górna in the Intra-Sudetic Basin, Poland.
1389 *Scientific reports*, 12(1), 1-26. <https://doi.org/10.1038/s41598-022-15644-2>.
- 1390 Putnis A, Hinrichs R, Putnis CV, Golla Schindler U, Collins LG (2007) Hematite in porous red-clouded
1391 feldspars; evidence of large-scale crustal fluid-rock interaction. *Lithos* 95: 10-18.
- 1392 Putnis CV, Ruiz-Aguido E (2013) The mineral-water interface: Where minerals react with the environment.
1393 *Elements* 9: 177-182.
- 1394 Ramseyer K, Boles JR, Lichtner PC (1992) Mechanism of plagioclase albitization. *Journal of Sedimentary*
1395 *Petrology* 62: 349-356.
- 1396 Reicherter K, Froitzheim N, Jarosiński M, Badura J, Franzke H-J, Hansen M, Hübscher H, Müller R, Poprawa P,
1397 Reinecker J, Stackebrandt W, Voigt T, Von Eynatten H, Zuchiewicz W (2008). Alpine tectonics north of
1398 the Alps. In: McCann (ed), *The Geology of Central Europe, Volume 2: Mesozoic and Cenozoic*.
1399 Geological Society, London, pp 1233-1285.
- 1400 Richter L., Diamond LW (2022) Characterization of hydrothermal fluids that alter the upper oceanic crust to
1401 spilitic and epidosite: Fluid inclusion evidence from the Semail (Oman) and Troodos (Cyprus) ophiolites.
1402 *Geochimica et cosmochimica acta*, 319: 220-253.
- 1403 Ricordel C, Parcerisa D, Thiry M, Moreau M-G, Gómez-Gras D (2007) Triassic magnetic overprints related to
1404 albitization in granites from the Morvan massif (France). *Palaeogeography Palaeoclimatology*
1405 *Palaeoecology* 251: 268-282. <https://doi.org/10.1016/j.palaeo.2007.04.001>.
- 1406 Ruffet G, Innocent C, Michard A, Féraud G, Beauvais A, Nahon D, Hamelin B (1996) A geochronological
1407 ⁴⁰Ar/³⁹Ar and ⁸⁷Rb/⁸⁷Sr study of K–Mn oxides from the weathering sequence of Azul, Brazil.
1408 *Geochimica et Cosmochimica Acta* 60: 2219–2232.
- 1409 Saigal GC, Morad S, Bjørlykke K, Egeberg PK, Aagaard P (1988) Diagenetic albitization of detrital K-feldspars
1410 in Jurassic, Lower Cretaceous, and Tertiary clastic reservoir rocks from offshore Norway, I Textures and
1411 origin. *Journal of Sedimentary Petrology* 58:1003-1013.
- 1412 Sandström B, Annersten H, Tullborg EL (2010) Fracture-related hydrothermal alteration of metagranitic rock
1413 and associated changes in mineralogy, geochemistry and degree of oxidation: a case study at Forsmark,
1414 central Sweden. *International Journal of Earth Science* 99: 1-25.
- 1415 Schmitt J-M (1986) Albitisation triasique, hydrothermalisme jurassique et altération supergène récente:
1416 métallogénie des gisements uranifères du Rouergue. *Doct. ès Sciences Thesis*, Strasbourg, Louis Pasteur
1417 University: p 240.
- 1418 Schmitt J-M (1992) Triassic albitization in southern France: an unusual mineralogical record from a major
1419 continental paleosurface. In: Schmitt JM, Gall Q (eds) *Mineralogical and geochemical records of*
1420 *paleoweathering*. ENSMP, Mémoires des Sciences de la Terre, Paris, pp 115–131
- 1421 Schmitt J-M (1994) Geochemical modeling and origin of the Triassic albitized regolith in Southern France. 14th
1422 *International Sedimentological Congress*, Aug. 20-28, Recife, Brazil, Abstracts S8, p. 19-21.

- 1423 Schmitt J-M, Baubron JC, Bonhomme MG (1984) Pétrographie et datations K–Ar des transformations minérales
1424 affectant le gîte uranifère de Bertholène (Aveyron-France). *Mineralium Deposita* 19:123–131.
- 1425 Schmitt J-M, Clement JY (1989) Triassic regolithization: a major stage of pre-enrichment in the formation of
1426 unconformity related deposits in Southern France. *Metallogeneses of uranium deposits*. IAEA Technical
1427 Committee Meeting, Vienna 542(8): 93-113.
- 1428 Simon K, Hoefs J (1987) Effects of meteoric water interaction on Hercynian granites from the Südschwarzwald,
1429 southwest Germany. *Chemical Geology* 61: 253-261.
- 1430 Słaby E, Götze J (2004) Feldspar crystallization under magma-mixing conditions shown by
1431 cathodoluminescence and geochemical modelling—a case study from the Karkonosze pluton (SW Poland).
1432 *Mineralogical Magazine*, 68(4), 561-577
- 1433 Słaby E, Martin H (2008) Mafic and felsic magma interaction in granites: the Hercynian Karkonosze pluton
1434 (Sudetes, Bohemian massif). *Journal of Petrology* 49: 353-391.
- 1435 Słaby E, Galbarczyk-Gąsiorowska E, Baszkiewicz A (2002) Mantled alkali-feldspar megacrysts from the
1436 marginal part of the Karkonosze granitoid massif (SW Poland). *Acta Geologica Polonica* 52: 501-519.
- 1437 Słaby E, Galbarczyk-Gąsiorowska E, Seltmann R, Müller A (2007) Alkali feldspar megacryst growth :
1438 Geochemical modelling. *Mineralogy and Petrology* 89: 1-29.
- 1439 Sobczyk A, Danišik M, Wojewoda J, Śliwiński W, Raczyński P, August C (2013) Zircon (U-Th)/He
1440 thermochronology of the Permian Weissliegendes sandstone in the Fore-Sudetic Homocline (SW Poland).
1441 *Mineralogia - Special Papers* 41: 81.
- 1442 Sobczyk A, Danišik M, Aleksandrowski P, Anczkiewicz A (2015) Post-Variscan cooling history of the central
1443 Western Sudetes (NE Bohemian Massif, Poland) constrained by apatite fission-track and zircon (U-
1444 Th)/He thermochronology. *Tectonophysics* 649: 47-57.
- 1445 Sobczyk A, Sobel ER, Georgieva V (2020) Meso-Cenozoic cooling and exhumation history of the Orlica-
1446 Śnieżnik Dome (Sudetes, NE Bohemian Massif, Central Europe): Insights from apatite fission-track
1447 thermochronometry. *Terra Nova* 32 (2): 122–133.
- 1448 Steinthorsson S, Helgason O (1992) Maghemite in Icelandic basalts. *Mineralogical Magazine* 56: 185-199.
- 1449 Stillings LL, Brantley SL (1995) Feldspar dissolution at 25°C and pH 3: Reaction stoichiometry and the effect of
1450 cations. *Geochimica et Cosmochimica Acta* 59: 1483-1496.
- 1451 Stober I, Bucher K (1999a) Origin of salinity of deep groundwater in crystalline rocks. *Terra Nov* 11:181–185.
1452 <https://doi.org/10.1046/j.1365-3121.1999.00241.x>
- 1453 Stober I, Bucher K (1999b) Deep groundwater in the crystalline basement of the Black Forest region. *Applied*
1454 *Geochemistry* 14: 237–254. [https://doi.org/10.1016/S0883-2927\(98\)00045-6](https://doi.org/10.1016/S0883-2927(98)00045-6)
- 1455 Sun SS, Eadington PJ (1987) Oxygen isotope evidence for the mixing of magmatic and meteoric waters during
1456 tin mineralization in the Mole Granite, New South Wales, Australia. *Economic Geology* 82: 43-52.

- 1457 Suzuki K, Adachi M (1994) Middle Precambrian detrital monazite and zircon from the hida gneiss on Oki-Dogo
1458 Island, Japan: their origin and implications for the correlation of basement gneiss of Southwest Japan and
1459 Korea. *Tectonophysics* 235: 277–292. [https://doi.org/10.1016/0040-1951\(94\)90198-8](https://doi.org/10.1016/0040-1951(94)90198-8)
- 1460 Suzuki K, Adachi M, Tanaka T (1991) Middle precambrian provenance of Jurassic sandstone in the Mino
1461 Terrane, central Japan: Th-U-total Pb evidence from an electron microprobe monazite study. *Sedimentary
1462 Geology* 75: 141–147. [https://doi.org/10.1016/0037-0738\(91\)90055-I](https://doi.org/10.1016/0037-0738(91)90055-I)
- 1463 Suzuki K, Adachi M, Kajizuka I (1994) Electron microprobe observations of Pb diffusion in metamorphosed
1464 detrital monazites. *Earth and Planetary Science Letters* 128: 391–405. [https://doi.org/10.1016/0012-1079-
1465 821X\(94\)90158-9](https://doi.org/10.1016/0012-1079-821X(94)90158-9)
- 1466 Taylor HP (1977) Water/rock interactions and the origin of H₂O in granitic batholiths. *Journal of Geological
1467 Society of London* 133: 509-558.
- 1468 Taylor HP (1977) Water/rock interactions and the origin of H₂O in granitic batholiths: Thirtieth William Smith
1469 lecture. *J Geol Soc London* 133:509–558. doi: 10.1144/gsjgs.133.6.0509
- 1470 Thiry M, Liron MN, Dubreucq P, Polton J-C (2017) *Curiosités géologiques du massif de Fontainebleau, Guide
1471 géologique*, BRGM Editions., 115 p.
- 1472 Thiry M, Milnes AR, Rayot V, Simon-Coinçon R (2006) Interpretation of palaeoweathering features and
1473 successive silicifications in the Tertiary regolith of Inland Australia. *Journal of the Geological Society*
1474 163: 723-736.
- 1475 Vasconcelos PM, Becker TA, Renne PR, Brimhall GH (1994) Direct dating of weathering phenomena by
1476 ⁴⁰Ar/³⁹Ar and K–Ar analysis of supergene K–Mn oxides. *Geochimica et Cosmochimica Acta* 58: 1635–
1477 1665.
- 1478 Vercruyse C (2011) Mise en évidence et datation de la paléosurface crétacée dans le massif des Vosges.
1479 Rapport No O110620CVER, Centre de Geosciences, Ecole des Mines de Paris, Fontainebleau, France, 75
1480 p.
- 1481 Waychunas GA (1991) Oxide Minerals: Petrologic and magnetic significance. *Reviews in Mineralogy* 25: 38-46.
- 1482 Wierzchołowski B (1976) Granitoidy kłodzko-złotostockie i ich kontaktowe oddziaływanie na skały osłony
1483 (studium petrograficzne). *Geologia Sudetica* 11: 7-147.
- 1484 Wilamowski A, (2002) Chloritization and polytypism of biotite in the Łomnica granite, Karkonosze Massif,
1485 Sudetes, Poland: Stable isotope evidence. *Chemical Geology* 182(2-4): 529-547.
- 1486 Williams ML, Jercinovic MJ, Goncalves P, Mahan K (2006) Format and philosophy for collecting, 1096
1487 compiling, and reporting microprobe monazite ages. *Chemical Geology* 225: 1–15. <https://doi.org/10.1016/j.mce.2005.09.009>
- 1489 Wojdyr M (2010) Fityk: A general-purpose peak fitting program. *Journal of Applied Crystallography* 43: 1126-
1490 1128.

- 1491 Yao KFE (2013) Albitization and oxidation of the granitoid rocks related to the Triassic paleosurface in the
1492 Sudetes (SW Poland). Thesis École Nationale Supérieure des Mines de Paris in Paris (France) and
1493 Państwowy Instytut Geologiczny - Państwowy Instytut Badawczy in Warsaw (Poland). 164 p.
1494 <http://pastel.archives-ouvertes.fr/pastel-00971314>
- 1495 Yerle J.-J., Thiry M (1979) Albitisations et minéralisations uranifères dans le socle et les sédiments permo-
1496 houillers du bassin de Brousse-Broquiès (Aveyron, France). Bulletin BRGM 4: 275-290.
- 1497 Žák J, Klominsky J (2007) Magmatic structures in the Krkonoše–Jizera Plutonic Complex, Bohemian Massif:
1498 evidence for localized multiphase flow and small-scale thermal–mechanical instabilities in a granitic
1499 magma chamber. *Journal of Volcanology and Geothermal Research* 164: 254-267.
- 1500 Žák J, Verner K, Sláma J, Kachlík V, Chlupáčová M (2013) Multistage magma emplacement and progressive
1501 strain accumulation in the shallow- level Krkonoše- Jizera plutonic complex, Bohemian Massif.
1502 *Tectonics* 32: 1493-1512.
- 1503 Zheng Y-F (1993) Calculation of oxygen isotope fractionation in anhydrous silicate minerals.
1504 *Geochim Cosmochim Acta* 57:1079–1091. doi: 10.1016/0016-7037(93)90042-U.
- 1505 Zijdeveld, JDA. (1967) A.C. demagnetisation of rocks: analysis of results. In: Collinson, DW, Creer,
1506 KM, Runcorn, SK (Eds.), *Methods in Paleomagnetism*. Elsevier, Amsterdam, pp. 254–286.
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1 **Albitization and oxidation of Variscan granitoid rocks related to the Post-Variscan paleosurface in the**
2 **Sudetes (Bohemian Massif, SW Poland)**

3
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20
21 **Abstract:**

22 The reddened granitoid facies in the basement of the Polish Sudetes exhibits two categories of alteration spatially
23 arranged with respect to fractures: (1) saussuritization and sericitization within light-colored facies in the interior
24 of fracture-bounded blocks and (2) albitization and hematitization in a reddened facies occurring adjacent to
25 fracture walls. These alterations are associated with chloritization of primary ferromagnesian minerals and the
26 development of secondary minerals such as quartz, K-feldspar, apatite, prehnite, calcite, and titanite. We link
27 these parageneses and the observed zonation to a unique alteration event consisting of an interplay of chemical
28 reactions of variable spatial extent. The complete albitization of the feldspars (plagioclase and K-feldspar)
29 adjacent to fractures points to a significant supply of Na for the albite neof ormation and availability of oxygen to
30 form the associated hematite. The dating of the iron oxides by paleomagnetism and the secondary monazite
31 associated with the albitized facies by U-Th-Pb_{total} unambiguously indicates their Post-Variscan ages. In this
32 context, the alterations are related to regionally widespread Post-Variscan paleosurface. Weathering profiles
33 formed in phreatic groundwater environments downgradient of highlands that provided a hydrological head.
34 Sodium supply was likely to origin from the gigantic salt playas that characterized Permian and Triassic

35 lowlands. Weathering was interrupted by an advance of the Mesozoic transgression. The recognition of these
36 weathering profiles over extensive parts of the Variscan Belt provides the basis for reconstructing the Post-
37 Variscan paleosurface and constraining Triassic and Post-Triassic geodynamics, including erosion rates and
38 geomorphological evolution of the Paleozoic massifs in Europe.

39

40 **Keywords:**

41 albitization, oxidation, granitoid, dating, paleosurface, Post-Variscan, Poland.

42

43 **Running title:**

44 Albitization related to the Post-Variscan paleosurface

45

46 **Statements and declarations:**

47 The authors declare that they have no known competing financial interests or personal relationships that
48 could have appeared to influence the work reported in this paper.

49

50 **Acknowledgements:**

51 This work is the result of a quest that was started more than 40 years ago at the Ecole des Mines de Paris. The
52 French team likes to acknowledge the two participating Polish Institutions for the organization of the field work
53 campaigns and for two scientific stays of K.Y. in the labs with access to CL. We acknowledge the IGP
54 Paleomagnetism group for access and help with 2G cryogenic magnetometer paleomagnetic and rock magnetic
55 measurements as well as the CLIMAG group at the LSCE for access to the rock magnetic measurements
56 performed in their laboratories. The authors thank Jean-Bernard Edel from the University of Strasbourg and an
57 anonymous reviewer for their thoughtful comments, which significantly strengthened the earlier version of the
58 article. Authors are also indebted to Tony Milnes from University of Adelaide for his discussions and language
59 edition.

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62

63 **INTRODUCTION**

64 In crystalline rocks, albitization is commonly considered to be a deep and high-temperature
65 metasomatic alteration linked to granite cooling during exhumation (e.g. Putnis et al. 2007; Engvik et al. 2008;
66 Sandström et al. 2010; Morad et al. 2010). Besides, in basaltic rocks, albitization with typical albite, chlorite, and
67 epidote assemblage is considered to be related to pervasive alteration by deep circulation of heated seawater
68 (spilitization) under relative lower temperature (Kelley et al. 1992; Alt et al. 2010; Richter and Diamond 2022).
69 Studies of albitization in granitic and other basement rocks have been mostly undertaken by petrologists who
70 usually consider its petrographical aspects and recognise the late-stage nature of the parageneses. However, the
71 geometric aspects of the albitization phenomenon have never been thoroughly investigated. In this paper we
72 describe the albitization as a result of superficial weathering, spatially and temporally associated with the Post-
73 Variscan paleosurface, by analogy with situations described locally in the French Massif Central (Yerle and
74 Thiry 1979; Schmitt 1986, 1992; Schmitt and Clement 1989; Parcerisa et al. 2009) and in NE Spain (Fàbrega et
75 al. 2019).

76 We focus our study on the Polish Sudetes where paleomagnetic dating of Post-Variscan
77 remagnetization overprints of the Paleozoic crystalline basement has already been well-reported in the literature
78 (e.g. Edel et al. 1997). Our aim is to sort out whether the albitization observed in the Sudetic granitoids could
79 have been of superficial (a weathering phenomenon) rather than of deep-seated hydrothermal origin. For this
80 purpose, we describe detailed mineralogical and petrographical features of the albitized facies rocks. Special
81 attention is paid to the petrographic evolution of the granitoids in relation to the reddened facies adjacent to
82 fractures. The relationship between primary and secondary mineral phases is emphasized with the focus to
83 establish a specific mineral paragenesis and geochemical conceptual model for the albitization, and to understand
84 its association with the contemporaneous formation of hematite. Subsequently, the results of paleomagnetic
85 dating of iron oxides and U-Th-Pb_{total} dating of monazites in the reddened facies are presented in order to explore
86 their respective links with the Post-Variscan paleosurface.

87 **GEOLOGICAL SETTING**

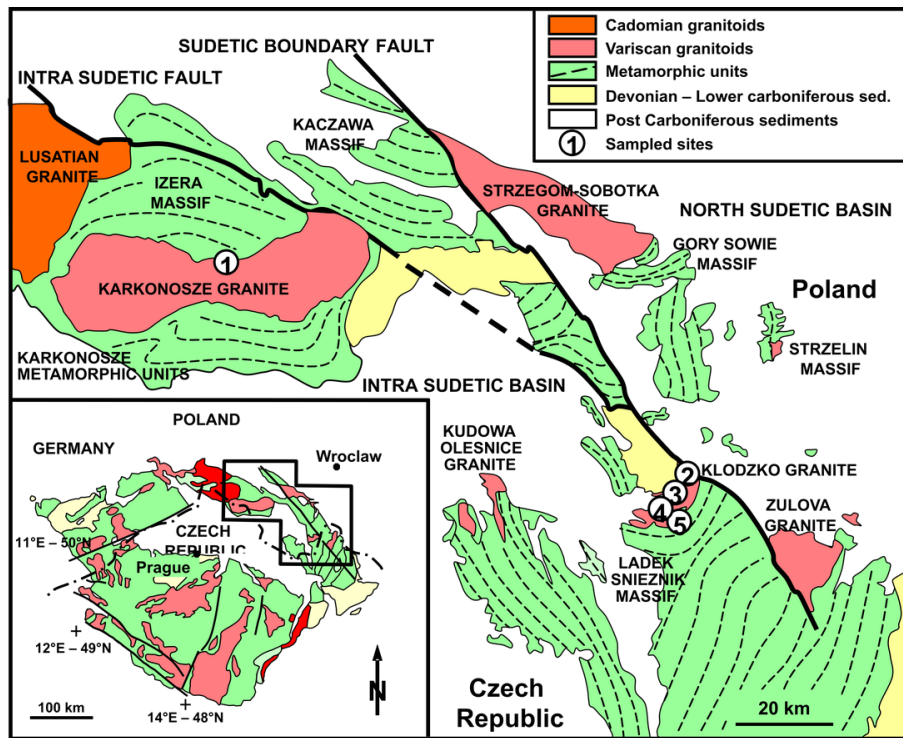
88 **Regional frame**

89 The mountain range of the Sudetes is the topographic expression of a regional-size fault block, about
90 300 km long and 50 to 80 km wide, at the NE margin of the Bohemian Massif (Fig. 1). The Sudetic block was
91 uplifted, together with the entire Bohemian Massif, in Late Cenozoic times along major NW-SE and NE-SW

92 trending faults at the foreland of the Eastern Alps and Carpathians (e.g. Reicherter et al. 2008), while to the NE it
93 is accompanied by the geologically similar Fore-Sudetic block that is currently not elevated.

94 The Sudetes occur on the northeastern periphery the internal zone of the Variscan belt of Europe, whose
95 mostly crystalline, Late Neoproterozoic to Carboniferous basement rock complexes are arranged in a mosaic of a
96 complex tectonic collage, amalgamated and deformed during Late Devonian through to Carboniferous Variscan
97 tectonic events (Matte et al. 1990; Aleksandrowski and Mazur 2002; Kroner et al. 2008). The collage is intruded
98 by syn- to late orogenic Carboniferous granitoid plutons, the most voluminous of which are those of Karkonosze,
99 Strzegom-Sobótka and Kłodzko - Złoty Stok (e.g. Mazur et al. 2006, 2007).

100 The basement rocks are locally overlain by syn- to late-tectonic Carboniferous intramontane basins and
101 remnants of a now much eroded post-orogenic Permo-Mesozoic sedimentary-volcanic cover (Fig. 1), which
102 during Late Cretaceous – Early Paleogene times underwent compressional deformation: gentle folding and
103 inversion-related faulting (Głuszyński and Aleksandrowski 2022). To the south, in the Intra-Sudetic Basin, Early
104 Permian red beds and shales with volcanics up to 1250 m thick are followed by Late Permian and Early Triassic
105 fluviatile deposits of Buntsandstein facies, which do not exceed a few tens of meters in thickness (e.g. Dziedzic
106 and Teisseyre 1990). To the north, in the North Sudetic Basin, there is a dominantly sedimentary series including
107 Early Permian red beds with volcanics and Late Permian Zechstein facies carbonates, sulphates and clastics,
108 averaging about 1000 m in thickness. The Early Triassic Buntsandstein of variegated sandstones, totalling about
109 600 m in thickness, are capped by Röt dolomites and later Keuper evaporites and Muschelkalk limestones (e.g.
110 Baranowski et al. 1990; Chrzastek and Wojewoda 2011). In south-western and western Poland and eastern and
111 northern Germany, the Early Triassic clastics that are accompanied locally by the Buntsandstein exceed 1000 m
112 in thickness (Feist-Burkhardt et al. 2008). The Early Permian and Early Triassic clastic deposits reflect intense
113 erosion and denudation of the southerly located Variscan internal massifs, including those of the Sudetes. The
114 Late Permian Zechstein carbonates and sulphates as well as the late-Early to Middle Triassic carbonates, on the
115 other hand, mark the episodes of marine transgression. The Late Permian transgression apparently left in the
116 Sudetes only a thin sedimentary sequence of tens of metres thick, whereas that of the late-Early to Middle
117 Triassic might have resulted in an originally much thicker sedimentary succession, similar to those in the
118 adjacent areas (cf. e.g. Feist-Burkhardt et al. 2008). This is suggested by increasingly deepening facies of the
119 now up to 100 m-thick Muschelkalk deposits towards their erosive top (Chrzastek and Wojewoda 2011). Also,
120 the original presence of a relatively thick and extensive cover of marine Jurassic sediments in the Sudetes cannot
121 be entirely ruled out, despite their present absence possibly due to pre-Cenomanian erosion.



122

123 **Fig. 1** Geological map of the Sudetes (modified after Aleksandrowski et al. 1997). (1) Szklarska Poręba Huta
 124 quarry; (2) Laski quarry; (3) Laski village; (4) Chwalisław valley; (5) Mt Kopciowa

125 **Sampling**

126 Our sampling was concentrated on the reddened granitic rocks in the quarry at Szklarska Poręba Huta
 127 on the Karkonosze pluton and in several exposures in the eastern part of the Kłodzko - Złoty Stok pluton, at
 128 Laski and in the Chwalisław valley (Fig. 1). The focus was put on large exposures of rocks with distinct signs
 129 of reddening that are typically bound to fractures. In this study we use the term fracture to any brittle
 130 discontinuities in a rock, whether or not they imply fracture-parallel or -perpendicular displacement, i.e. various
 131 types of both joints and faults. Non-reddened rocks were also sampled in order to characterise differences
 132 between the reddened and non-reddened facies.

133 *Karkonosze granitoid massif*

134 The late-orogenic Karkonosze granitoid massif in the western part of the Sudetes is composed of a
 135 number of granitic facies thought to correspond to fractional crystallization, mixing with mafic magmas and
 136 hybridization processes (Žák and Klomínský 2007; Slaby and Martin 2008). The pluton resulted from sequential
 137 emplacement over a brief period of time at ca. 320-310 Ma (Machowiak and Armstrong 2007; Žák et al. 2013;
 138 Kryza et al. 2012, 2014; Kusiak et al. 2014).

139 Sampling of the Karkonosze pluton was undertaken in the Szklarska Poręba Huta quarry (N50°49'39'',
 140 E15°29'38'') on the NW margin of the Karkonosze massif, close to its contact with the metamorphic envelope

141 (Fig. 1). Samples were obtained from a steep northern wall of the quarry where there was an exposure of two
142 rock types in sharp intrusive contact: (1) a fine-grained equigranular leucocratic granite (so called aplogranite)
143 and (2) a medium- to coarse-grained porphyritic biotite granite. Both granites contain thick (up to 25 cm) aplite
144 veins and centimetre-wide quartz veins. According to Słaby et al. (2007), the granite from Szklarska Poręba Huta
145 belongs to the geochemically furthest evolved granite types in the Karkonosze composite pluton.

146 The porphyritic granite from the Szklarska Poręba Huta quarry contains large phenocrysts (up to 10 cm)
147 of perthitic K-feldspar (about 65-99% Or) mantled by plagioclase (about $Ab \geq 83\%$) and enclosed in medium-
148 grained groundmass (Borkowska 1966; Słaby et al. 2002). Relative to the porphyritic granite, the aplogranite is
149 enriched in quartz and plagioclase and impoverished in biotite and alkali-feldspar (Kozłowski et al. 1975).

150 *The Kłodzko - Złoty Stok pluton*

151 The Kłodzko - Złoty Stok (KZS) pluton, the largest composite intrusion in central Sudetes (Fig. 1), is
152 composed off strongly diversified high-K metaluminous rocks, rich in mafic components (e.g. Lorenc 1994;
153 Bachliński and Bagiński 2007; Jokubauskas et al. 2014). The small magma batches were emplaced between
154 $349.0 \pm 3.7\text{Ma}$ and $331.5 \pm 2.6\text{Ma}$. Short-lived late Carboniferous rejuvenation of magmatic activity along major
155 fault zones caused intrusion of the the Laski monzogranite at $303.3 \pm 3.8/4.2\text{Ma}$ (Jokubauskas et al. 2018 and
156 references therein).

157 Exposures of reddened granitic rocks were sampled in the Laski quarry (N50°28'25" E16°48'34"),
158 Laski village (N50°27'40" E16°48'12") and a small old quarry near the creek of Mąkolnica (Chwalisław valley;
159 N50°25'54" E16°49'42"). Jointly were sampled non reddened, relatively dark-grey medium-grained granodiorite
160 with relatively uniform plagioclase composition (36-46% An) typical of the Laski area (Wierzchołowski 1976).
161 The quarry in the Chwalisław valley exposes granodiorite intruded by dykes (up to about. 1 m wide) of
162 spessartite. Both rocks had been previously dated, yielding ages of $336.7 \pm 2.3\text{Ma}$ and $321.1 \pm 3.1\text{Ma}$,
163 respectively (Mikulski et al. 2013). Non-reddened granodiorite is a grey, medium- to fine-grained, equigranular
164 rock with relative anorthite-rich plagioclase (39-44% An), whereas the fine-grained microporphyritic spessartite
165 is composed of more sodic plagioclase (Wierzchołowski 1976; Awdankiewicz and Awdankiewicz 2010;
166 Mikulski et al. 2013).

167 **ANALYTICAL METHODS.**

168 Powder X-ray diffraction (XRD) analyses were carried out at Mines-Paris, using a PANalytical X'Pert
169 system fitted with a Cu-anode and X'celerator detector, on fragments of larger K-feldspar crystals separated by
170 hand-picking from the bulk rock samples. XRD scans were run from 2 to 60° (2 θ). The XRD diagrams were
171 interpreted using the EVA© Bruker software. In the case of overlapping (131) and (1-31) peaks, deconvolution

172 was undertaken using the FITYK software (Wojdyr 2010) with Pearson VII function applied to fit the peak
173 shapes. Al-Si ordering of the K-feldspar structure was evaluated on the basis of the triclinicity parameter
174 according to the method of Goldsmith and Laves (1954).

175 Selected thin sections from the reddened rock samples were studied using conventional petrographical
176 microscopy and cathodoluminescence (CL) techniques at the Institute of Geological Sciences, University of
177 Wrocław. The CL study made use of a CL MK3 cold cathode gun mounted on a polarizing optical microscope.

178 Systematic sampling for paleomagnetic analysis was performed focused on the reddened facies at four
179 principal outcrops, three of them situated in the Kłodzko – Złoty Stok Massif (Laski quarry, Laski village,
180 Chwalisław valley) and one in Szklarska Poręba Huta area. The samples for paleomagnetic analyses were drilled
181 in the vicinity of principal fractures. The samples were collected as oriented drill cores using a petrol-driven
182 hand drilling machine and as block samples oriented in field by measuring dip and dip azimuth.

183 They were cut into standard 2.5 x 2.5 cm cylinders for paleomagnetic experiments. Thermal
184 demagnetization was subsequently performed on the cylinders either in the paleomagnetic laboratory of the
185 Institute du Physique du Globe de Paris using a vertical three-axial 2G cryogenic magnetometer and at the
186 Institute of Geophysics of the Polish Academy of Sciences in Warsaw using a 2G Enterprise SQUID
187 magnetometer. Stepwise demagnetization of the NRM was applied on the samples, starting at room-temperature,
188 and if possible, embracing the complete temperature range up to the hematite temperature interval around 700°C.
189 Alternatively, some selected samples were demagnetized using the alternating field technique at the Institute of
190 Geophysics of the Polish Academy of Sciences in Warsaw. The resulting data are systematically corrected for
191 the geographic position of each individual sample.

192 Further data processing was performed using either the PaleoMac5 software (Cogné 2003) or the
193 PALMAG software provided by Lewandoski et al. (1997) Characteristic paleomagnetic directions were
194 determined by a least-square fit of linear segments (Kirschvink, 1980) in orthogonal Zijderveld diagrams
195 (Zijderveld 1967; see also supplementary material 2, Fig. S1 to S3), either on equal areas or in great circle fits.
196 The PaleoMac5 and PALMAG software offer the opportunity to choose the grouped directions and to calculate
197 respective site-mean directions and paleomagnetic pole statistics (Fischer, 1953; Mc Fadden and McElhinny
198 1988). The analyses of thermal demagnetization of NRM curves and thermomagnetic experiments
199 (magnetization and saturation remanence during heating curves) allow the identification of magnetic carriers of
200 the paleomagnetic directions (for instrumental details and rock magnetic results please see also supplementary
201 material 2, Fig. S4 and S5).

202 The chemical compositions of feldspars and micas were analysed by EPMA (Electron Probe Micro
203 Analysis) at the Scientific and Technological Centers of the Universitat de Barcelona (CCiTUB), using Cameca

204 SX-51 and JEOL JXA-8230 electron microprobes, both operating at 15 kV and 1.5 nA. The CL images of the
205 samples were used to guide the placing of the analyses. The table of EPMA results is included in the
206 supplementary material 1.

207 Monazite dating was carried out using the EPMA U-Th-Pb_{total} method following the recommendations
208 included in Cocherie et al. (1998), Cocherie and Albarede (2001), and Williams et al. (2006). Monazite grains
209 suitable for EPMA analysis were found by means of BSE imaging using a Hitachi TM-1000 electron microscope
210 at the Manresa School of Engineering (Spain). U-Th-Pb_{total} EPMA analyses were undertaken at the Scientific and
211 Technological Centers of the Universitat de Barcelona (Spain) using a Jeol JXA-8230 electron microprobe
212 operating at 15 kV, 40 nA of current intensity and a beam diameter of 1 μm . An in-house reference monazite
213 from the Petaca District (Mexico), with an age of $1332 \pm 5 \text{ Ma}$ (2σ), was analyzed as a control standard jointly
214 with the monazite samples. The age of each analysed monazite was calculated using the MonaziteAge software
215 based on the works of Suzuki and Adachi (1991, 1994) and Suzuki et al. (1994) and which is included with the
216 Jeol JXA-8230 instrument. The initial acquired data were prepared for subsequent statistical age analysis by
217 eliminating the analyses which presented oxide totals higher and lower than $100 \pm 5\%$. Subsequently, age
218 outliers were discarded using a stepwise process based on box-plot graphs. The final ages of the different
219 monazite generations were determined by deconvolution of the ages histogram using the Microsoft Excel add-in
220 ISOPLOT 4.15 from the Berkeley Geochronology Center (Ludwig 2003). A total of 189 EPMA analyses were
221 made and 147 analyses (77% of the total) were retained including 18 from the Szklarska Poręba Huta quarry and
222 129 from the Chwalisław valley section. The table of monazite EPMA results is included in the Electronic
223 Supplementary material 1.

224 The oxygen isotope composition of plagioclase and K-feldspar was determined by Secondary Ion Mass
225 Spectrometry (SIMS) using a CAMECA IMS 1280HR ion microprobe at the Centre de Recherches
226 Pétrographiques et Géochimiques (CRPG), Vandœuvre-lès-Nancy, France. Samples were mounted in 25 mm
227 diameter epoxy mounts and the surface was metallographically polished. Afterwards the samples surfaces were
228 imaged under optical cathodoluminescence to differentiate textural domains within the crystals. The detailed
229 analytical conditions during the SIMS session are described in Fàbrega et al. (2017). The granite samples were
230 analysed jointly with in-house plagioclase and K-feldspar standards. The analyses were placed in homogeneous
231 luminescent domains in crack- and inclusion-free areas of the crystals. A total of 28 and 31 analyses were carried
232 out on plagioclase and K-feldspar grains, respectively. Afterwards, the craters of the SIMS analyses were
233 observed under BSE-imaging and analyzed by EPMA to check that the craters were homogeneously excavated
234 and determine the chemical composition linked to each isotopic analysis. The instrumental mass fractionation
235 (IMF) of the raw results was corrected by means of response surface methodology (RSM) following the
236 procedures and using the response surface models described in detail by Fàbrega et al. (2017). The SIMS results

237 including the instrumental setting, the raw and IMF-corrected $\delta^{18}\text{O}$ values, and the CL-images of the samples are
238 included in the Electronic Supplementary Information. All the $\delta^{18}\text{O}$ values of this paper are relative to VSMOW.

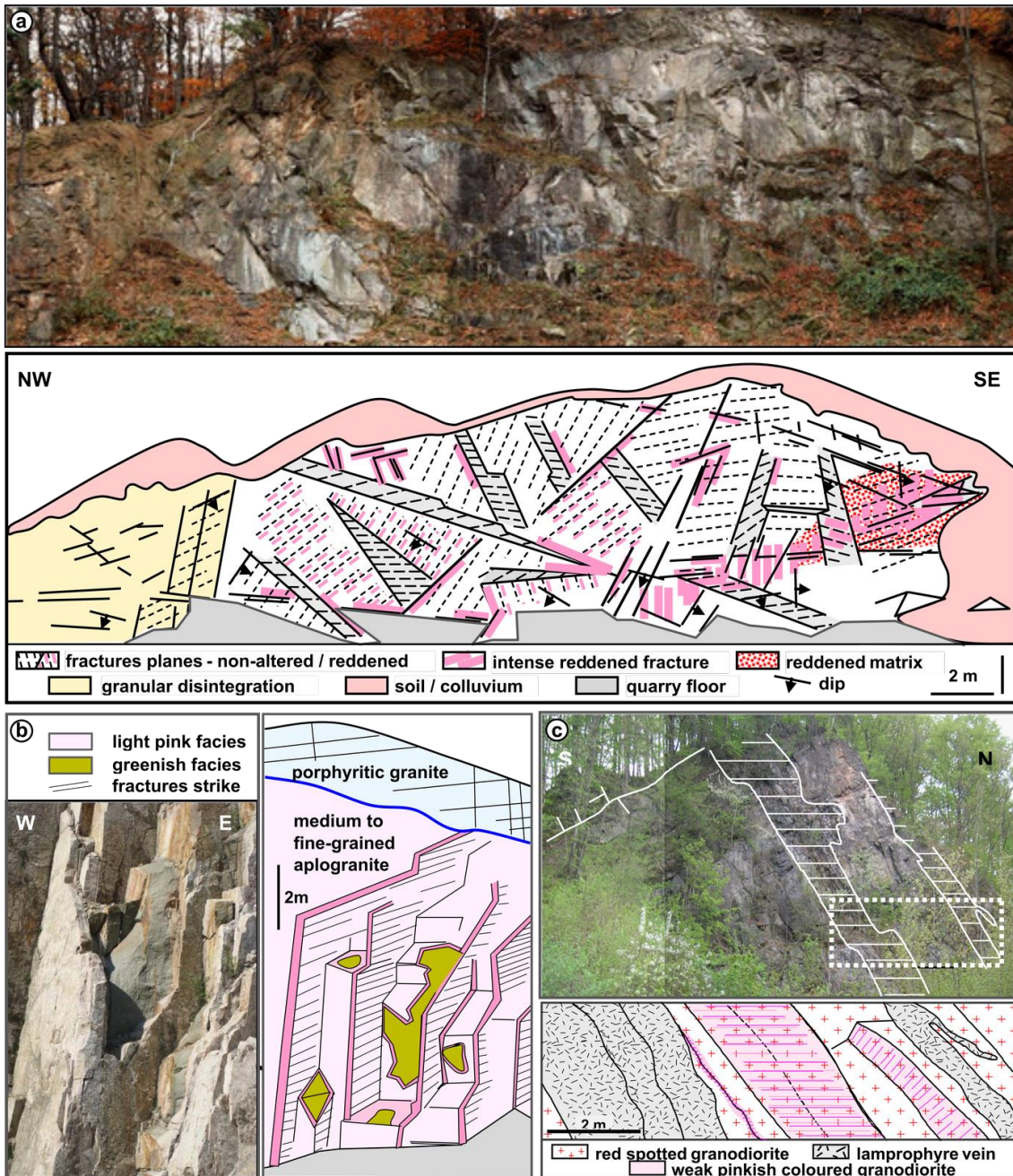
239 SPATIAL DISTRIBUTION OF THE REDDENING

240 A process of granite reddening has resulted in both a pervasive coloration homogeneously distributed
241 within the rock volume at some area and a more localised development of red to pinkish granite facies allied to
242 the fracture network at other areas. The most typical and widespread type of alteration is a **fracture controlled**
243 **reddening** with reddish/pinkish coloration restricted to narrow bands along fracture planes. The granodiorite of
244 the Laski quarry and the aplogranite of the Szklarska Poręba Huta quarry are typical examples for this. The most
245 intense alteration is a **pervasive and homogeneous reddening** with the entire rock impregnated by the reddish
246 pigments, and fracture planes coated with dark red iron oxides. The Laski village section is typical of this
247 pervasive reddened facies. The less intense alteration is a **red spotted facies** with pinkish spots distributed
248 within non-reddened rock. The granite of Chwalisław valley outcrop is typical of this weakly altered facies.

249 In the Laski quarry the fractures are metre-spaced, cutting the rock into well-defined 0.5 to 1 m³ blocks
250 (Fig. 2a). The extent of fracture controlled reddening is closely related to the rank of the fracture, i.e. it is weak
251 along minor fractures and more intense along major ones. The reddening penetrates the rock as deep as 1-10 cm.
252 The coloration is most intense along fracture walls but fades further away from the fractures. There is also a
253 direct relation between the frequency of the fractures and the intensity of coloration. In the SE part of the quarry,
254 the most intensive reddened facies occur along a weakly defined metre-wide zone, characterized by particularly
255 dense fracturing. Similarly, the porphyritic granodiorite exposed on Mt Kopciowa has fracture controlled
256 reddening restricted to weak, narrow staining fronts, where its large K-feldspar phenocrysts are partly or entirely
257 pink along the fractures.

258 In the Szklarska Poręba Huta quarry, the porphyritic granite is generally not reddened. However, the
259 fine to medium-grained aplogranite exhibits a light pink reddening along steep N40° major and N135° related
260 fractures, which penetrates the rock to a distance of 5-15 cm to 30-40 cm away from the fractures. Where
261 fractures are sparse, the aplogranite is greenish in the center of the blocks (Fig. 2b) but in densely fractured
262 (“chopping fractures”) zones the rock is pervasively pink.

263 The Chwalisław valley granodiorite outcrop is a typical example of the red spotted facies (Fig. 2c). A
264 weak fracture controlled pinkish coloration has developed in narrow zones along fractures but further away from
265 the fracture planes there is a scatter of spots together with small pinkish patches in the non-reddened rock. There
266 is no obvious direct relationship between lamprophyre dikes and the reddening of the granodiorite. In fact the
267 lamprophyre also has a faint pinkish coloration along the fracture planes, but their observance is hindered by the
268 dark color of the rock.



269

270 **Fig. 2** Spatial arrangement of the reddening. (a) Photo and sketch of the fracture-controlled reddening in the
 271 Laski quarry. The highly fractured zones in the SE of the quarry are the most reddened. (b) Szklarska Poręba
 272 Huta quarry facies. The reddening proceeds along the steep fractures that cut the fine-grained aplogranite. (c)
 273 Sketch of the reddish alteration facies in the Chwalisław valley outcrop. The alteration is parallel to the
 274 lamprophyre veins that intruded the granodiorite

275 **MICROSCOPIC ASPECTS OF REDDENING**

276 Optical and cathodoluminescence petrography reveal some systematic differences between the non-
 277 reddened and reddened rock facies. We present petrographical comparison of the affected mineralogical

278 constituents, but those that are neither affected by nor related to the staining phenomenon are omitted from the
279 comparison.

280 **Plagioclase**

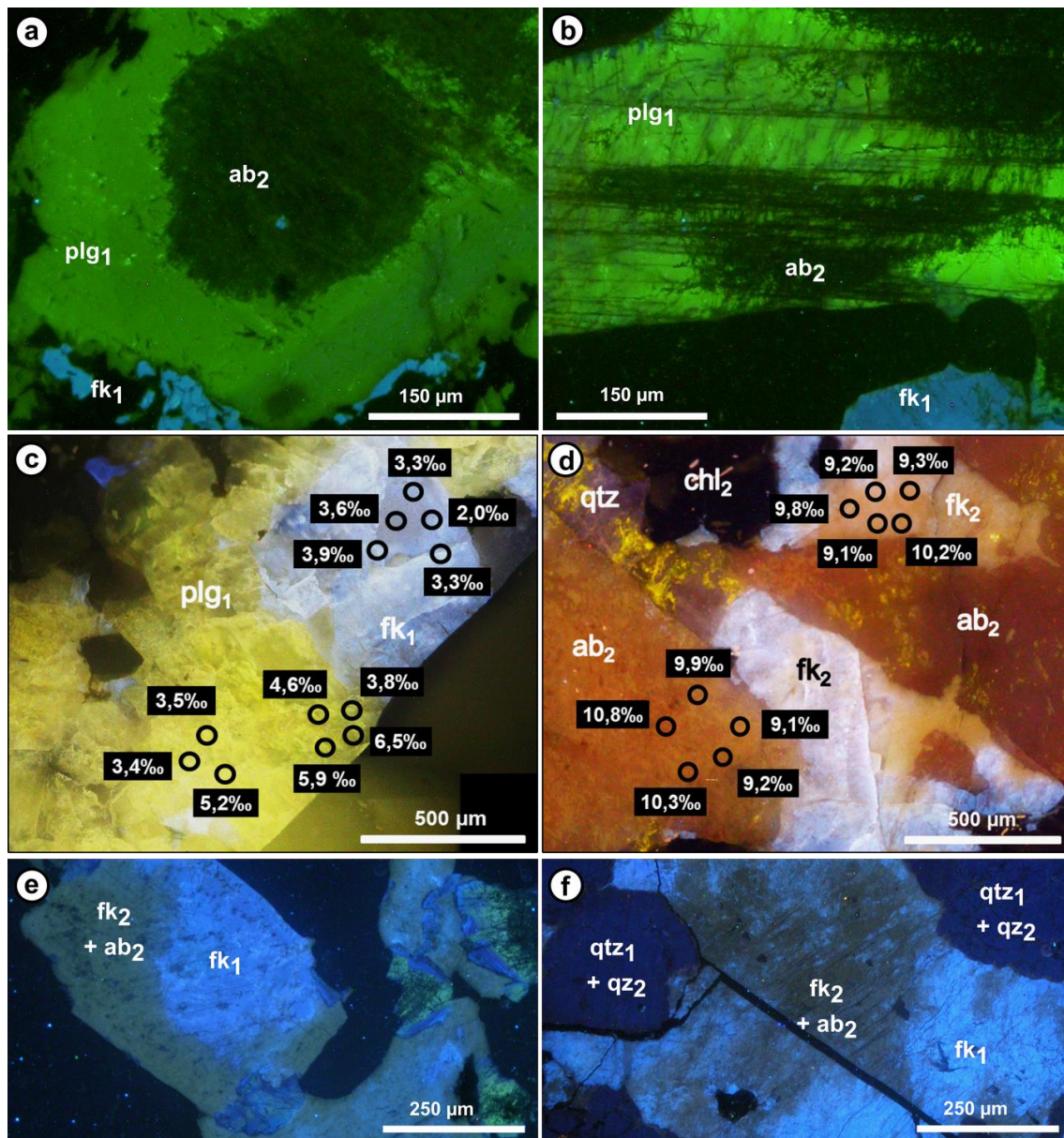
281 In non-reddened granites plagioclase appears white to the naked eye. Large crystals are often normal-
282 zoned with Ca-enriched cores and Ca-poor margins: some crystals are also rimmed with nearly pure albite.
283 Polysynthetic twinning is common. It also occurs as inclusions in the K-feldspar phenocrysts (in porphyritic rock
284 types) and constitutes most of the rock groundmass. The luminescence is typical for igneous plagioclase with
285 yellow-green CL (Fig. 3a and 3b), that turns blue in the albitic rims. Regions with increased intensity of yellow-
286 green CL correlates with a higher Ca content (cf. Slaby et al. 2002; Slaby and Götze 2004). Partial sericitization
287 of primary plagioclase is often observed, especially in the Ca-enriched cores.

288 In reddened granites plagioclase appears pink to the naked eyes, due to hematite pigments which confer
289 a red cloudiness to spots where primary plagioclase has been thoroughly altered to secondary untwined albite
290 (i.e. albitization). While the hematite-free domains retain their primary CL characteristics, the pigmented
291 plagioclase is non-luminescent (Fig. 3a and 3b), which is typical of neogenic feldspars formed at low
292 temperatures (e.g. Marshall 1988; Götze et al. 2000; Leichmann et al. 2003; Powolny et al. 2022). Contrasting
293 CL conveniently mark out the geometrical relationship between primary plagioclase and secondary albite: Ca-
294 enriched crystal cores may be albitized leaving the outer Na-enriched zones unaffected (Fig. 3a) or secondary
295 albite may be associated with microcracks, twinning and/or cleavage planes (Fig. 3b).

296 Primary plagioclase compositions range from An₂₆ to An₈₄, and An₁₄ to An₅₁, in the Chwalisław biotite-
297 hornblende granodiorite and Laski valley granodiorite, respectively (Table 1). The oxygen isotope composition
298 of the primary plagioclase, determined in the biotite-hornblende granodiorite of Chwalisław, shows $\delta^{18}\text{O}$ values
299 ranging from +3 to +7 ‰ (Fig. 3c and Table 2), with no apparent correlation with the composition of plagioclase
300 determined by EPMA in the SIMS points, that ranges from An₃₆ to An₄₃.

301 Secondary albite shows a widespread micro porosity and a typical composition about An_{0.2},
302 independently of the type of parent rock and of the primary plagioclase composition (Table 1). In crystals where
303 the completeness of albitization has not been totally reached, a slightly higher Ca content was observed, with
304 An% values ranging from 6 to 14% (Table 1). The oxygen isotope composition of the secondary albite,
305 determined in the Chwalisław sector, is significantly heavier than the primary plagioclase, with $\delta^{18}\text{O}$ values
306 ranging from +7 to 11‰ (Fig. 3d and Table 2).

307



308

309 **Fig. 3** CL Petrography and oxygen isotope composition of albitized plagioclase and K-feldspar. (a) Plagioclase
 310 with a non-luminescent albitized core and yellow luminescent rim. (b) Albitization along the major twinning
 311 planes and micro-fractures. (c) $\delta^{18}\text{O}$ values of yellow luminescent primary plagioclase and blue luminescent
 312 primary K-feldspar and (d) non-luminescent secondary albite and turbid microclinized K-feldspar. (e)
 313 Cathodoluminescence images of partially albitized K-feldspar showing non-luminescent areas containing
 314 secondary microcline, patch perthites and quartz patches on the crystal rims. (f) Albitization of K-feldspar along
 315 microfractures. plg₁: primary plagioclase; ab₂: secondary stained albite; fk₁: primary K-feldspar; fk₂: secondary
 316 K-feldspar; qtz₁: primary quartz; qtz₂: secondary quartz

317

318 **Table 1** Mole % of feldspar components of plagioclase and K-feldspar

site	rock	plagioclase		K-feldspar	
		primary / albitized	An %	primary / microclinized	Or %
<i>Reddened facies</i>					
Szklarska Poręba	porphyritic granite	primary	1 - 3	microclinized	94 - 98
Laski Quarry	granodiorite	albitized	0 - 2	microclinized	95 - 98
<i>Spotted facies</i>					
Szklarska Poręba	porphyritic granite	primary	8 - 14	partially microclinized	78 - 93
Laski Quarry	granodiorite	albitized	4 - 6	partially microclinized	79 - 92
<i>Fresh rock</i>					
Laski Valley	granodiorite	primary	21 - 59	primary	85 - 91
Chwalislaw Valley	hornblende-biotite granodiorite	albitized	6 - 14	primary	72 - 93

* EPMA analyses converted to mole percent of the feldspar components by normalizing to eight oxygen atoms per formula unit.

319

320 **Table 2** $\delta^{18}\text{O}$ ‰ results of plagioclase and K-feldspar

site	rock	plagioclase		K-feldspar	$\delta^{18}\text{O}$ ‰
		primary / albitized	$\delta^{18}\text{O}$ ‰		
<i>Reddened facies</i>					
Chwalislaw Valley	hornblende-biotite granodiorite	albitized	6 - 11	microclinized	5 - 11
<i>Fresh rock</i>					
Chwalislaw Valley	hornblende-biotite granodiorite	primary	3 - 7	primary	2 - 4

321

322 Apart from hematite pigments, neogenic albite is sometimes associated with calcite and prehnite.
 323 Secondary albite is not twinned - its development erased the polysynthetic twinning in the primary plagioclase.
 324 This is in contrast to “classical” saussuritization in which the petrographic features of the primary plagioclase are
 325 preserved, at least partially (de la Roche 1957; Sandström et al. 2010; Jenkin et
 326 al. 1992; Morad et al. 2010).

327 **K-feldspar**

328 In non-reddened granites, K-feldspar is white to grey and often
 329 Carlsbad-twinned with varying amounts of microperthitic albite. In the
 330 porphyritic granitoids it is present as large (several centimeters)
 331 phenocrysts and as a constituent of the groundmass (<< 1 cm). Cross-hatch
 332 twinning is not a common feature but, if present, is developed in domains
 333 rather than whole crystals.

334 As observed using CL, any alteration of primary crystals is
 335 manifested by a change in the blue luminescence typical of igneous K-

336 feldspar (cf. Słaby et al. 2002; Słaby and Götze 2004; Powolny et al.
337 2022). The hematite-stained parts of K-feldspar crystals correspond to
338 regions of quenched, sometimes dark-brownish luminescence, similar to the
339 previously shown secondary albite (Fig. 3e and 3f). Such CL characteristics
340 match secondary feldspars formed in low-temperature diagenetic and
341 supergenic environments (e.g. Götze et al. 1999; González-Acebrón et al.
342 2010; González-Acebrón et al. 2012).

343 There is a clear relationship between hematite staining and
344 albitization of K-feldspar. Semi-quantitative measurements of the intensity
345 of XRD reflections from the two feldspar species reveal that a low and
346 rather constant albite content in K-feldspar phenocrysts in non-reddened
347 granitoids increases and becomes highly variable in the reddened facies
348 (Yao 2013). Primary light-colored K-feldspar phenocrysts are predominantly
349 formed of orthoclase but an increase in albite content is generally marked
350 by its transformation (recrystallization) to microcline, as indicated by an
351 increasing triclinicity. Since triclinicity reflects the state of Al-Si
352 ordering of K-feldspar structure it is likely that the increase of Al-Si
353 ordering, albitization and hematite-staining of the K-feldspar are
354 interrelated processes.

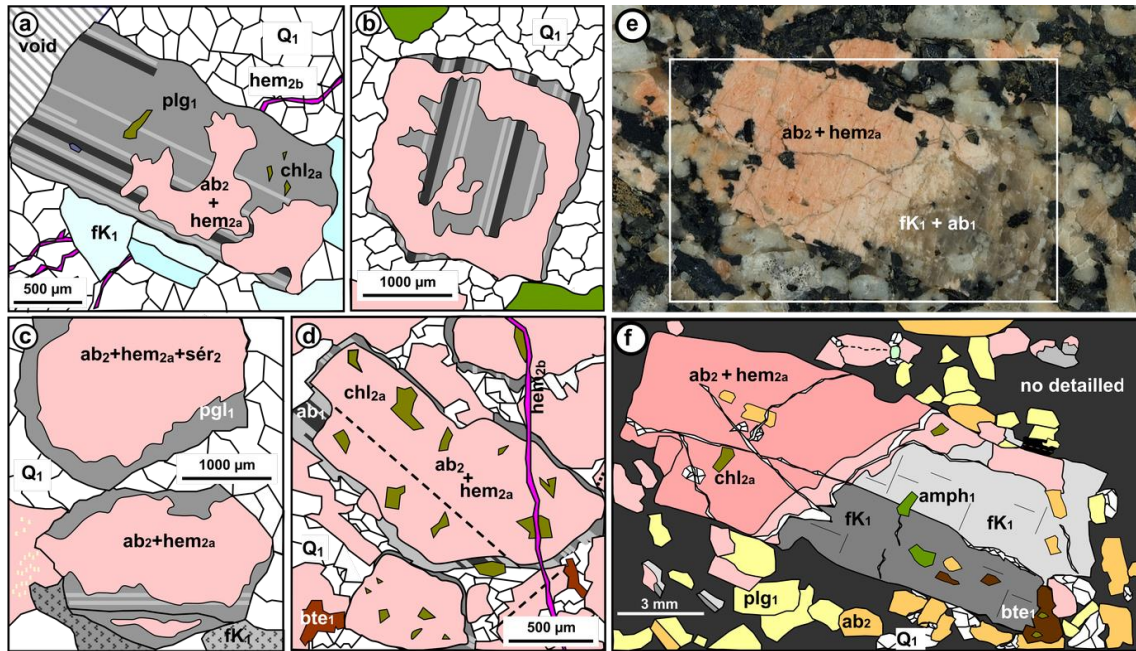
355 The composition of the primary K-feldspar ranged from Or₇₂ to Or₉₃ in
356 the the Chwalisław sector and from Or₈₅ to Or₉₁ in the Laski valley (Table 1).
357 The oxygen isotope composition of the primary K-feldspar (Table 2),
358 determined in crystals from the biotite-hornblende granodiorite of
359 Chwalislaw valley, present $\delta^{18}\text{O}$ values ranging from +2 to +4 ‰. The $\delta^{18}\text{O}$
360 values show a positive trend versus the increase in K⁺ content of the K-
361 feldspar (see Electronic Supplementary Information). In the quenched
362 luminescence regions, the primary orthoclase is replaced by secondary
363 microcline showing abundant micro porosity, with typical composition
364 varying from Or₉₄ to Or₉₈, independently of the parent rock and the
365 composition of the primary K-feldspar. The microclinization process is
366 accompanied by the development of patch perthites of highly pure albite,
367 with typical composition of An₀. The oxygen isotope composition of the
368 microclinized K-feldspar (Table 2), also determined in the biotite-
369 hornblende granodiorite of Chwalislaw, presents $\delta^{18}\text{O}$ values, ranging from +5

370 to +11 % and also shows a positive trend versus the orthoclase content (Or
371 %) of the K-feldspar.

372 **Albitization propagation**

373 Albitization of primary plagioclase occurred along microcracks and
374 cleavage planes (Fig. 4a). The albitized zones are patchy and amoeboid,
375 with no relation to the crystallographic directions in the primary
376 plagioclase. Often albitization developed from the margins towards the
377 central parts of a crystal, leaving the core as well as the primary albitic
378 outer rim unaffected (Fig. 4b). A common albitization pattern is observed
379 in some of the compositionally zoned plagioclases in which the Ca-enriched
380 core has been altered but the Ca-poor rims remains unaffected (Fig. 4c).
381 The albitized core often contains scattered sericite which corresponds to
382 an earlier phenomenon than albitization as it is also observed in non-
383 albitized plagioclase from the non reddened facies. In most altered facies,
384 primary plagioclase is perversavely albitized exept outer rims composed of
385 primary albite (Fig. 4d). These altered facies are intensely reddened and
386 often show micro-fractures filled with iron oxides.

387 Albitization of primary K-feldspar proceeds in a similar way as in
388 palgioclases: from the margins of the crystals along microcracks, twinning
389 and/or cleavage planes. In the reddened granitoids, euhedral to sub-
390 euhedral K-feldspar phenocrysts are partly or entirely pink (Fig. 4e and
391 4f). There is a direct correlation between the intensity of staining in the
392 granitoids and the amount and distribution of hematite pigmenting
393 phenocrysts and smaller K-feldspar crystals.



394

395 **Fig. 4** Sketches of albitized red/pinkish plagioclase and K-feldspar. (a)
 396 Plagioclase albitization along microfractures. (b) Irregular albitization
 397 of the outer rim of the primary plagioclase: the primary albitic rim
 398 remains unaffected. (c) Albitization of the seritized core of the primary
 399 plagioclase. (d) Complete albitization of the primary plagioclase crystal
 400 except the outer rim which is composed of primary albite. (e) Polished
 401 section of a partially reddened orthoclase phenocryst. The upper part of
 402 the phenocryst is pink and dull whereas the lower part is translucent. (f)
 403 Sketch of (e). The orthoclase phenocryst exhibits classic Carlsbad twinning
 404 in the translucent part and has inclusions of hornblende, chlorite,
 405 biotite, and plagioclase. In the pinkish part the twinning is erased, and
 406 the orthoclase is albitized and hematized. fk₁: primary K-feldspar; plg₁:
 407 primary plagioclase; Q₁: primary quartz; bte₁: biotite; amph₁: amphibole;
 408 ab₂: untwined secondary albite; chl_{2a}: secondary chlorite after biotite;
 409 hem_{2a}: pigmentary hematite; hem_{2b}: coarser-crystalline hematite

410 Fe-Mg minerals

411 Light to dark brown biotite is the main mafic mineral and minor
 412 greenish hornblende is commonly present in samples from the Kłodzko - Złoty
 413 Stok massif. In non-reddened samples both biotite and hornblende show signs
 414 of chloritization, though hornblende appears more resistant to this
 415 alteration.

416 Chloritization is significantly more intensive in the reddened
417 granitoids where all stages of the replacement process, from initial weak
418 alteration along the edges and cleavage planes of the primary crystals to
419 the formation of complete chlorite pseudomorphs, could be observed. Biotite
420 inclusions in hematite-stained plagioclase are invariably replaced by
421 chlorite. There is no clear evidence for a volume change during the
422 chloritization of the primary Fe-Mg minerals. Secondary microcrystalline
423 and flaky chlorite also occurs in microfractures and intergranular
424 boundaries. Pyrite, Ti-oxides and titanite are often associated with
425 secondary chlorite.

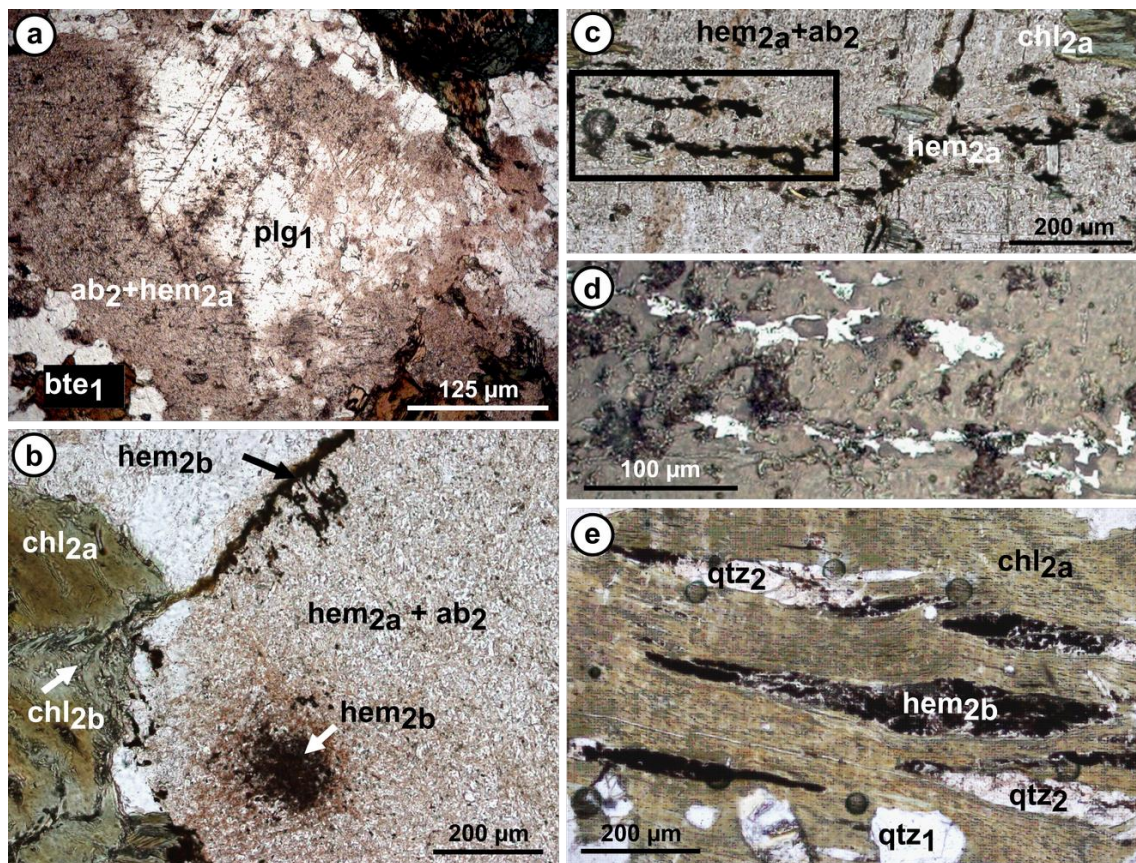
426 **Fe-minerals**

427 Fe-sulphides and oxides in the granitoids include pyrite, hematite,
428 maghemite, and magnetite. Pyrite occurs irrespective of hematite
429 pigmentation as relatively large (50-200 μm) euhedral crystals and lens-
430 shaped aggregates in the cleavages of chloritized biotite and hornblende.
431 Late anhedral grains fill also microfractures. The margins of most of the
432 observed pyrite grains are replaced by hematite.

433 Pigmentary hematite is responsible for red cloudiness observed in
434 plagioclase from the reddened granitoid facies under microscope. In the red
435 spotted facies plagioclase exhibits coexisting hematite-pigmented and
436 hematite-free domains which are primary igneous plagioclase (Fig. 5a) and
437 in pervasively reddened facies and in fracture controlled reddened facies
438 the plagioclase is entirely hematite-stained (Fig. 5b). Dusty hematitic
439 pigment is confined to albitized domains in plagioclase and K-feldspar.
440 Some observations suggest that the hematitic pigments are not directly
441 related to microporosity in the altered feldspars. Firstly, the pigment is
442 absent from the porous saussuritized feldspars of the non-reddened
443 granitoid facies. Secondly, it is not concentrated along the albitization
444 front which is enriched in micropores that relate to the volume-loss
445 replacement of the primary feldspar by secondary albite. Instead, hematite
446 is distributed more or less homogeneously through all of the albitized
447 feldspar. Microscopic flaky aggregates and subhedral larger grains of
448 hematite are often observed in intergranular boundaries located close to

449 chloritized biotite and hornblende (Fig. 5b) and in microfractures
450 (Fig. 5c) and, as well as in the cleavage of some chloritized biotite
451 (Fig. 5e), sometimes together with secondary quartz and/or calcite.

452 Maghemite and magnetite could not be recognized unequivocally under
453 the microscope nor by XRD analyses but their presence was unmistakably
454 detected by thermal demagnetization experiments (paleomagnetic experiments,
455 this study). In contrast to hematite, magnetite is restricted to the non-
456 reddened facies. Maghemite, on the other hand, was detected mainly in the
457 weakly reddened facies. Thus, maghemite may be regarded as a 'transitional'
458 Fe-mineral between the magnetite-dominated non reddened granitoids and
459 hematite-dominated reddened facies.



460
461 **Fig. 5** Hematite occurrences within albitized granitoids. a) Light pink
462 albitized plagioclase away from reddened fracture walls with light colored
463 relict of primary plagioclase. (b) Plagioclase entirely reddened and
464 albitized from pervasively reddened facies; hematite appears as pigments
465 within the secondary albite and as granules within microfractures. (c) Fine
466 red hematite pigments and elongated hematite flakes within secondary

467 albite; plane polarized. (d) Enlargement of (c); reflected light. (e)
468 Hematite lenses within the cleavage of a chloritized biotite crystal,
469 notice lenses of secondary quartz in similar position; plane polarized.
470 plg₁: primary plagioclase; ab₂: secondary albite; bte₁: primary biotite;
471 chl_{2a}: secondary chlorite after biotite; chl_{2b}: neogenic chlorite in micro-
472 fracture; hem_{2a}: secondary hematite pigment; hem_{2b}: secondary granular and
473 flaky subhedral hematite

474 **Other minerals**

475 Quartz of the non-reddened facies is essentially unaltered,
476 presenting a homogeneous dull-luminescence. Differently, the quartz grains
477 of the reddened facies are significantly altered, presenting a network of
478 micron sized non-luminescent veins of secondary quartz displaying in a
479 cobweb texture. Internal zoning of quartz have already been described in
480 the chloritized facies of the Sudetic Karkonosze granite (Wilamowski 2002),
481 and similar quartz alteration has been observed in the red albitized
482 facies of NE Spain (Fàbrega et al. 2019).

483 Sericite is more abundant in non-reddened samples and in weakly-
484 reddened rock away from fractures. It occurs mostly in plagioclase and
485 emphasizes the compositional zoning as it has preferentially developed in
486 the Ca-enriched plagioclase cores but is entirely absent from the albitic
487 rims. Sericitization obliterates the primary polysynthetic twinning in the
488 plagioclase host. Tiny sericite crystals (5 µm) occur in plagioclase in the
489 aplogranite of Szklarska Poręba Huta and less fractured granodiorites of
490 the Kłodzko - Złoty Stok massif. This form of sericite is associated with
491 secondary albite but not with Ca-bearing minerals such as prehnite and
492 epidote. However, larger sericite flakes (10-20 µm) formed at the expense
493 of Ca-enriched plagioclase in the reddened granodiorite from Chwalisław and
494 Laski are associated with secondary prehnite, albite, and sometimes also
495 calcite. Sericite is rare in K-feldspars.

496 Prehnite and less commonly calcite are either present as secondary
497 minerals associated with saussuritization of plagioclase and chloritization
498 of Fe-Mg minerals or as late primary phases healing fractures. While
499 calcite is observed exclusively in the reddened rock facies, disseminated

500 prehnite is more abundant in the non-reddened granitoids. Secondary
501 prehnite, apart from prehnite-sericite-albite aggregates in the
502 saussuritized Ca-enriched plagioclase cores, occurs also as monomineralic
503 fan- and bowtie-shaped aggregates in intergranular boundaries and as
504 elongated lenses in biotite. The latter are concordant with cleavage and
505 the inflation of the lenses disturbs the parallel alignment of the mica
506 flakes. They are observed only in partially chloritized biotite in the non-
507 reddened granitoids but are absent in the proximity of reddened fracture
508 walls. Similar lenses of secondary calcite are present in chloritized
509 biotite and hornblende in the reddened rock facies. Both prehnite and
510 calcite, sometimes accompanied by laumontite, are also observed as late
511 primary minerals healing microfractures cutting across the mineral
512 framework of the reddened granitoids.

513 Magmatic apatite is commonly present as small inclusions in biotite
514 and hornblende or as inherited inclusions in their chlorite pseudomorphs.
515 Nonetheless, in some chlorite pseudomorphs, we also observe apatite
516 inclusions that were considerably larger and at the same time less abundant
517 than typical apatite inclusions in unaltered primary Fe-Mg minerals. They
518 are likely a second generation of apatite resulting from recrystallization
519 and also dissolution of smaller crystals during chloritization, such as
520 those shown in the albitized facies in the Morvan (Parcerisa et al., 2009)
521 and in NE Spain (Fàbrega et al., 2019) and those linked to low-temperature
522 spilitic albitization in the Intra-Sudetic Basin (Powolny et al. 2022).

523 REE-bearing minerals, mostly epidote and monazite, are typically
524 subhedral to euhedral crystals around 50-200 μm in size, in the non-
525 reddened. The REE-rich epidote is more abundant in the reddened facies than
526 in the unaltered rocks and occurs as larger grains up to 200-400 μm .
527 Monazite in the reddened facies occurs in two different forms depending on
528 its position with respect to the reaction front. The crystals in a cm-scale
529 zone inside the reddish/pinkish facies in contact with the albitization
530 reaction front are mostly pseudomorphed by secondary monazite. This
531 secondary monazite has corroded grain boundaries and micron-sized porosity,
532 suggesting that the replacement occurs via a dissolution-reprecipitation

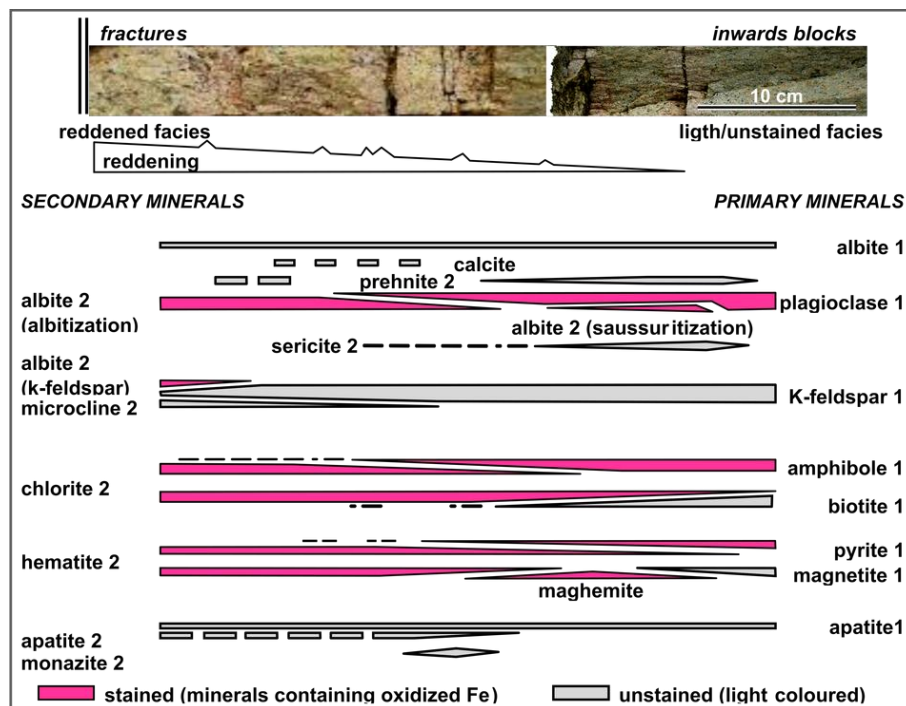
533 process. In contrast, in strongly reddened rocks, monazite grains can be
534 pseudomorphed by secondary apatite.

535 **Summary of petrographic observations**

536 Petrographic analyses of the reddened and non-reddened granitoids
537 provide a basis for discerning major differences in the mineralogical
538 composition of the two granitoid facies. Apart from hematite, which is
539 responsible for the red pigmentation, and quartz, the content of which is
540 fairly constant, the relative abundances of major and some minor
541 constituents vary with increasing reddening (Fig 6). In particular, the
542 amount of Ca-enriched plagioclase decreases largely at the expense of
543 secondary albite and is practically undetectable in the most intensely
544 reddened sample. While plagioclase in the non-reddened rocks is partially
545 sericitized, the plagioclase in the reddened facies is predominantly
546 albitized. To some extent K-feldspar follows the same pattern, with
547 albitization and microclinization of primary K-feldspar in the reddened
548 facies. Weak chloritization of biotite and amphibole is observed already in
549 the non-reddened granitoids but the process is dramatically more intense
550 with increasing reddening and leads to a complete disappearance of the
551 primary Fe-Mg minerals in the most reddened facies. Chloritization of
552 biotite, in particular where the alteration is intense, is often coupled
553 with the formation of secondary titanite and Fe-minerals along cleavage
554 planes and microfractures. Similarly, chloritized amphibole often contains
555 calcite along cleavages, indicating the removal of Ca from the chloritized
556 crystal.

557 Small amounts of prehnite occur in the unpigmented granitoids but it
558 becomes more abundant with reddening. However, its presence also depends on
559 the initial rock type as it is absent from the aplogranite of Szklarska
560 Poręba Huta and the granodiorite of Laski village, irrespective of the
561 extent of staining. The sericite content seems to be unaffected by
562 reddening. Dispersed calcite and laumontite are detected exclusively in the
563 reddened facies, though they are present as late fracture fillings in both
564 stained and unstained granitoids.

565 The presence of hematite pigments in feldspars is directly related to
 566 albitized domains. For instance, we have not observed albitized plagioclase
 567 that is hematite-free. Also, hematite pigments in K-feldspars were not
 568 observed in non-albitized regions. The crystallization of hematite pigments
 569 seems to be synchronous with albitization. If hematite formed later in
 570 micropores within secondary albite, it is expected that the migration of
 571 the hematitizing fluids followed some pathways. No such possible pathways
 572 were observed. Thus, assuming that hematite and secondary albite are
 573 contemporaneous, the albitization must have taken place under oxidizing
 574 conditions. Furthermore, secondary minerals related to albitization near
 575 the fractures are stained with iron oxides whereas secondary
 576 saussuritization minerals, corresponding to albite-sericite-prehnite
 577 aggregates, are not affected by the reddening. Albitization-
 578 saussuritization alterations grade progressively from fractures inwards
 579 into the fracture blocks; they do not overlap or cut across each other as
 580 expected if they were distinct unrelated alteration stages.



581
 582 **Fig. 6** Spatial distribution of mineral parageneses in relation to marginal
 583 fractures and core of granitoid fracture blocks

584 **AGE DATING AND GEODYNAMICS**

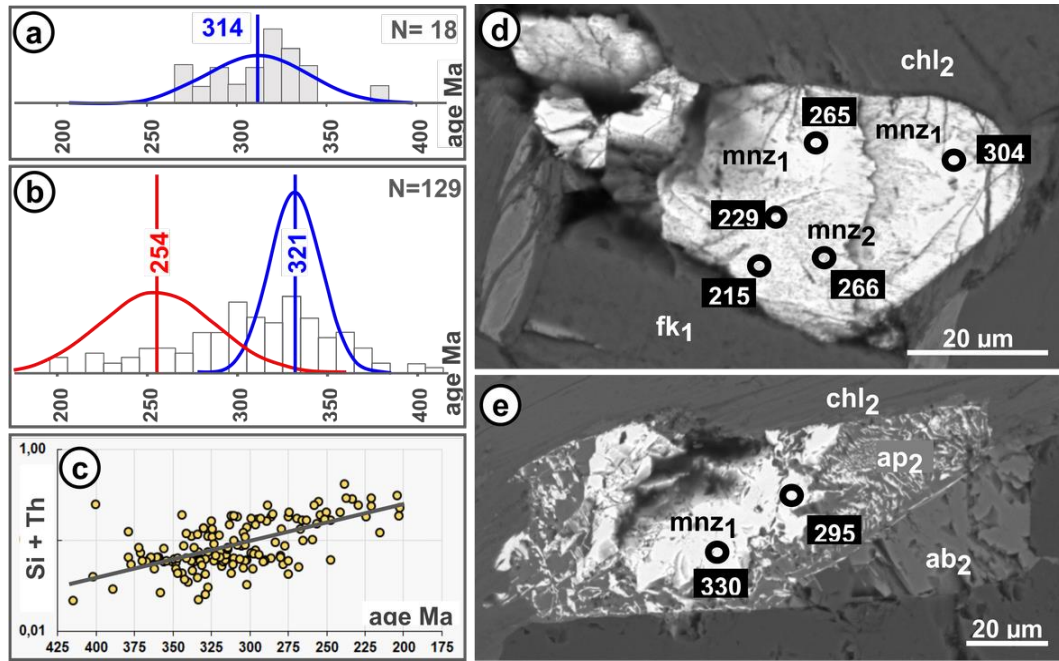
585 Dating of the albitization provides a clear tool for determining if
586 it could be related to deep, high-temperature metasomatic processes linked
587 to granite cooling or alternatively a product of weathering on uplift and
588 exposure of the granitoid mass. We compared the results of two independent
589 methods of dating: on one hand paleomagnetic dating of iron minerals and on
590 the other hand U-Th-Pb_{total} radiochronology of monazite, a phosphate able to
591 incorporate high U and Th concentrations.

592 **Monazite dating**

593 The Szklarska Poręba Huta quarry section provided only one sample
594 containing mainly unaltered monazite. Deconvolution of the monazite U-Th-
595 Pb_{total} ages histogram (Fig. 7a) provides an average age of 314 ± 56 Ma (2σ).
596 Although we cannot discard the possibility that some of these points could
597 represent a mix of primary and secondary mineral ages. The
598 Chwalisław valley section provided two samples containing numerous monazite
599 grains for U-Th-Pb_{total} radiochronology. The deconvolution of the monazite U-
600 Th-Pb_{total} ages histogram (Fig. 7b) yields two distinct age peaks, one at
601 254 ± 24 Ma (2σ) and a second at 321 ± 8 Ma (2σ) with a relative misfit of
602 0.988.

603 The textural analyses of samples from the Chwalisław valley section
604 show that the older ages occur in well preserved grains from the non-
605 reddened facies as well as in primary domains preserved within altered
606 grains of the slightly reddened/pinkish facies. On the other hand, the
607 younger ages (centered around 254 Ma) are almost observed in the monazite
608 grains with dissolution-reprecipitation features in the reddish/pinkish
609 facies close to the reaction front (Fig. 7d and 7e). The resetting of the
610 monazite geochronometer is accompanied with compositional changes,
611 including an increase in the Huttonite component [Th,Si(PO₄)], Fe and Al,
612 and a loss of REE+Y (Fig. 7c). These compositional changes in relationship
613 with younger ages, and the systematic occurrence of the younger ages in the
614 more altered monazite grains close to the reaction front strongly support
615 that primary monazite in the reddish/pinkish albitized facies were variably
616 recrystallized at about 254 Ma. The broad distribution observed in the 254

617 Ma peak relative to the 321 Ma peak is caused by less younger age data (in
 618 comparison with older age data).



619
 620 **Fig. 7** Dating of monazite. (a) Mean age obtained from the histogram of ages
 621 of samples from the Szklarska Poręba Huta quarry, sample WL 006. (b)
 622 Deconvolution ages obtained from the histogram of ages of samples from the
 623 Chwalisław valley section, samples WL 018 and WL 018a. (c) Single U-Th-
 624 Pb_{total} age of a monazite grain in the pinkish facies near the albitization
 625 front showing primary monazite (clear patches) partially replaced by
 626 microporous secondary monazite, BSE image. (d) Single U-Th- Pb_{total} age of a
 627 monazite grain in the strongly reddened rocks partially replaced by
 628 secondary apatite, BSE image. mnz₁: primary monazite; mnz₂: secondary
 629 monazite; ap₂: secondary apatite; ch₂: chloritized biotite; fk₁: primary K-
 630 feldspar, ab₂: secondary albite

631 **Paleomagnetic dating**

632 Albitization and reddening caused by the presence of hematite, are
 633 shown to be closely associated as they reveal no spatial discrepancy at
 634 outcrop or microscope scale. We thus consider albitization and hematization
 635 to be cogenetic. The dating of the hematite formation is possible by dating
 636 the paleomagnetic secondary remagnetization signal. This would then allow
 637 to date the albitization event.

638 The different components of the paleomagnetic signal recorded within
639 the analyzed oriented samples have been separated by stepwise thermal
640 demagnetization experiments and are related to their specific magnetic
641 carriers (see also supplementary material 2). According to the respective
642 temperature interval at which a specific decay in the magnetic intensity
643 curve was observed, one or a succession of several magnetic carriers were
644 identified in the sample. In principle, thermomagnetic and microscopic
645 analysis showed three different types of magnetic mineral assemblages were
646 observed within the analyzed samples: (1) hematite dominated in pervasive
647 reddened facies and in fracture controlled reddened facies; (2) coexisting
648 hematite-maghemite mostly in fracture controlled reddened facies further
649 away from the fractures; and (3) maghemite dominated mainly in spotted
650 facies.

651 In addition, maghemite-bearing mineral assemblages sometimes register
652 residual, presumably primary magnetite. This was mainly observed in samples
653 distant from fracture zones and in the less reddened spotted facies. As
654 well, a weak contribution of goethite was sometimes observed which was
655 demagnetized after application the 120°C demagnetization step. Goethite is
656 considered as negligible component compared to the other magnetic carrier
657 minerals and is most likely linked to a weak recent overprint by surface
658 weathering of the outcropping rock sections.

659 The stepwise thermal demagnetization allowed us to obtain the vector
660 directions recorded by the magnetic carrier minerals at the moment of their
661 formation. These directions were considered to be characteristic for either
662 the maghemite or hematite components, or both, if their inclination and
663 declination values were stable according to the respective considered
664 temperature intervals. It is thus possible to sometimes extract two
665 slightly different directional intervals from one single sample if the two
666 main carriers diverge in their mean direction. The results were 'sub-
667 grouped' according to their directional information.

668 The extracted directional information was then transformed into
669 magnetic paleopoles (e.g. Butler 1998) that were subsequently projected
670 onto the apparent polar wander path (APWP) reference curve for Eurasia

671 (Edel and Düringer 1997; Besse and Courtillot 2003). In total 275
672 individual samples were analyzed for the calculation of the mean paleopoles
673 carried by hematite or hematite-maghemite.

674 The severe rejection criteria for paleomagnetic samples in our
675 dataset are mainly due to high error defined by values of $\alpha 95$ (Kirschvink
676 1980; Fisher et al. 1987). Additionally, we eliminated samples that did not
677 show any linear segments of the demagnetization vector on a Zijderveld
678 diagram or a clear stable endpoint near the origin of a Zijderveld plot, as
679 described by McFadden and McElhinny (1988). The resulting 18 averaged
680 paleopoles all show acceptable $\alpha 95$ values (< 15) according to the
681 recommendations of e.g. Fisher et al. (1987) and very often plot directly
682 on or close to the Apparent Polar Wander Path (APWP; Table 3). Six
683 paleopoles were calculated for one or two directions only. For these
684 paleopoles the mean angle deviation (MAD) values are given instead. All
685 poles were listed in the Table 3. The paleomagnetic data set also includes
686 additional paleopoles that correspond to the primary Devonian magnetization
687 of the granitoid rocks, carried by primary magnetite (not shown here).

688

689 **Table 3** Results from paleomagnetic data analyses, corresponding to the
690 averaged site-wise paleopoles plotted in Fig. 8. MAD = mean angle
691 deviation; TH = thermal demagnetization; AF = alternating field
692 demagnetization; mgt = magnetite; mhm = maghemite; hm = hematite;
693 goe = goethite; LC = low coercivity compounds; HC = high coercivity
694 compounds.

paleo- poles	site outcrop	age	Nb of samples	α 95	MAD	type of analysis	pole latitude	pole longitude	magnetic carriers
1	Sklaska Poreba	Late Carboniferous	3	8.9		TH	20	125	mgt, mhm, hm
2	Chwalisław	Late Carboniferous	2		10.1	AF	23	126	LC+HC
3	Laski quarry	Late Carboniferous	3	8.7		TH	27	141	mgt, mhm, hm
4	Laski quarry	Middle Permian	4	8.7		TH	41	182	mhm, hm
5	Laski quarry	Late Permian	2		6.7	TH	53	161	goe*, hm
6	Laski village	Late Permian	1		3.8	TH	55	170	hm
7	Sklaska Poreba	Late Permian	3	10.1		TH	47	157	mhm
8	Laski quarry	Late Triassic	4	5.1		AF	54	127	LC+HC
9	Laski quarry	Triassic/Jurassic	2		7.7	AF	51	107	HC
10	Chwalisław	Triassic/Jurassic	3	3.3		AF + TH	49	118	mhm
11	Chwalisław	Triassic/Jurassic	3	7.3		TH	42	99	mhm, hm, goe*
12	Sklaska Poreba	Triassic/Jurassic	3	7.2		TH	36	104	mhm, hm
13	Laski quarry	Early Jurassic	4	9.1		TH	53	74	mhm
14	Chwalisław	Early Jurassic	8	5.9		TH	58	83	mhm, goe*
15	Laski quarry	Early Jurassic	2		6.6	AF + TH	73	93	mhm, hm
16	Sklaska Poreba	Early Jurassic	3	5.9		AF	74	67	LC+HC
17	Laski village	Middle Jurassic	6	5.0		TH	73	170	hm
18	Laski village	Middle Jurassic	1		10.4	TH	79	114	hm

695

696 Figure 08 shows the mean paleopoles for all four studied outcrops
697 including Laski quarry, Laski village, Chwalisław valley, and Szklarska
698 Poręba Huta. Four main groups of paleopoles and some 'isolated' average
699 paleopoles plot along the APWP curve for Eurasia

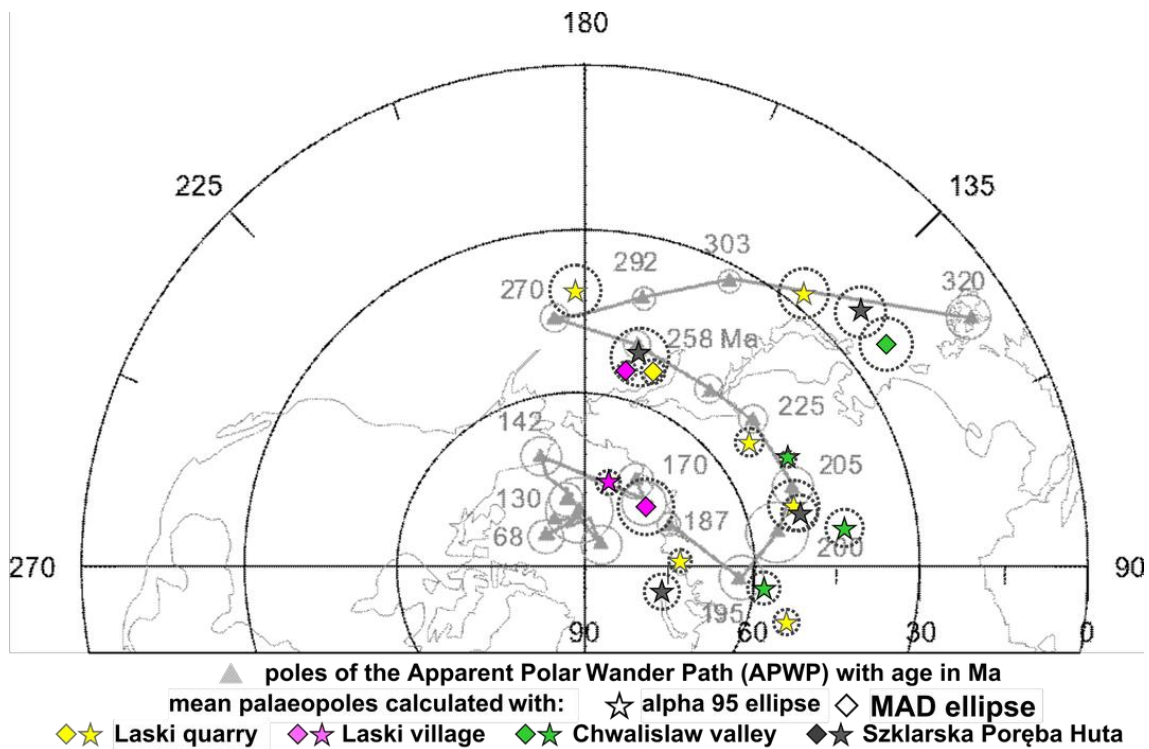
700 (1) Samples from paleopoles 1, 2, and 3 have a mixed magnetic assemblage
701 which is dominated by primary magnetite. These represent the oldest
702 group with Late Carboniferous ages (between 320 to 303 Ma) and are
703 represented in all outcrops (with exception of Laski Village, that
704 exhibits exclusively pervasive reddened facies) whichpresumably
705 correspond to the primary age of the granitoid. All other paleopoles
706 record secondary remagnetization ages.

707 (2) An 'isolated' average paleopole of Middle Permian age near the 270 Ma
708 reference point from Laski quarry (paleopole 4).

709 (3) A cluster of Late Permian ages close to 258 Ma from the Szklarska
710 Poręba Huta quarry, Laski quarry and the basal part of the Laski village
711 section (paleopoles 5, 6, 7).

712 (4) A relatively 'isolated' average paleopole of Late Triassic age close
713 to 225 Ma from Laski quarry (paleopole 8).

- 714 (5) A group of ages near Triassic/Jurassic boundary (around 205 Ma)
715 comprising four sets of paleopoles from the Laski quarry (paleopole 9),
716 Chwalisław valley (paleopoles 10 and 11) and Szklarska Poręba Huta
717 (paleopole 12) samples.
- 718 (6) A group of paleopoles of Middle Jurassic ages (between 175-165 Ma;
719 paleopoles 17 and 18). This group is clearly different because it is
720 solely carried by samples from the upper part of the Laski village
721 section that is pervasively reddened/hematized. The intense
722 remagnetization of these samples seems to have erased the primary
723 magnetization carried by magnetite and the potential Permo-Triassic
724 overprint. The lower part of the section still retains the Permo-
725 Triassic signal.
- 726 (7) A group of Early Jurassic paleopoles (13, 14, 15, 16) plots between
727 the 195 to 187 Ma reference point, but they show a significant S to SW
728 offset from the APWP. Part of this directional offset may be due to the
729 fact that these samples (from the Laski quarry, Chwalisław valley, and
730 Szklarska Poręba Huta) may have been remagnetized successively multiple
731 times during the Permo-Triassic and later again during Early Jurassic
732 times. If so, the hematite in these samples could carry a mixed
733 cumulative age signal that is probably less well extractable than the
734 initial remagnetization from primary magnetite to maghemite and/or
735 hematite during the Permo-Triassic alteration phase.



736

737 **Fig. 8** Stereographic projection of the paleomagnetic results for the
 738 calculated average paleopoles for the study sites. The paleopoles are
 739 compared to the APWP for Eurasia (grey line) using the combined curve of
 740 Edel and Düringer (1997) from the Permian to Middle Jurassic (grey circles)
 741 and Besse and Courtillot (2003) from the Middle Jurassic until present
 742 (grey triangles). Star symbols represent mean poles with their respective
 743 α 95 confidence circle, diamond symbols are paleopoles with their
 744 respective MAD confidence circle, corresponding to Table 3

745 In our petrographic studies magnetite was solely found in the non-
 746 reddened facies whereas the magnetite detected by thermal demagnetization
 747 in samples of the weakly reddened facies probably relates to residual tiny
 748 crystals included in quartz. Maghemite is mainly detected in the weakly to
 749 intermediately pigmented facies distant from the reddened fractures at
 750 Laski and Szklarska Poręba Huta and in red spotted facies from Chwalisław
 751 valley. According to the literature, maghemite is metastable with respect
 752 to hematite and is considered to have formed by the slow oxidation of
 753 magnetite under weak oxidative conditions (Steinthorsson and Helgason 1992;
 754 Bowles et al. 2011). Hematite is dominant in the pervasively reddened
 755 facies in the Laski village section and is indicative of strong oxidative
 756 environments.

757 A chronology appears between the different magnetic facies of
758 alteration. The less reddened spotted facies which is mainly dominated by
759 maghemite carries the oldest remagnetization ages while the pervasively
760 reddened and albitized facies shows the youngest ages. In the mean time,
761 incipient reddening and albitization appears as the less intensely altered
762 facies and conversely the pervasively reddened and even pervasively
763 albitized without primary plagioclase and K-feldspar relict appears as the
764 most intensely altered facies. This conjunction and the fracture controlled
765 reddening that diminish obviously from most intensely altered facies towards
766 the less altered spotted facies led us to consider that these various
767 gradients agree with a succession of facies from the base to the top of the
768 alteration profiles. This suggests also that, with time, the active
769 alteration front progressively retreated upwards into shallower weathering
770 horizons.

771 It is of particular note that the Late Permian U-Th-Pb_{total} age at
772 254 ± 24 Ma obtained for secondary monazites in samples from the
773 Chwalisław valley section corresponds to the paleomagnetic poles 10, 11,
774 and 14 near 258 Ma on the Eurasian APWP (Fig. 8). These paleomagnetic
775 poles, were obtained for weakly pigmented facies from Lasky quarry,
776 Sklarska Proeba Huta, and Laski village, from samples taken further away
777 from reddened fractures that contain significant amounts of maghemite. On
778 one hand, maghemite developed in the early stages of albitization, under
779 less oxidative conditions. On the other hand, secondary monazite formed in
780 the immediate vicinity of the albitization front and was rapidly altered
781 away from this front. The consistency in independent ages of the albitizing
782 and oxidizing environments obtained from completely different secondary
783 minerals and by means of distinctly different dating methods provide
784 confidence in datation results. This does not contradict with the younger
785 paleomagnetic ages obtained in this study for which we do not have access
786 to separately datable minerals.

787 **Links to other age determinations**

788 The older age obtained for monazite from the Chwalisław valley
789 section (321 ± 8 Ma) is within the range of the previously published primary

790 crystallization ages for the Kłodzko - Złoty Stok (KZS) Pluton between
791 349.0±3.7 Ma and 331.5±2.6 Ma (U-Pb, zircon; Jakubauskas et al. 2018 and
792 references therein). The younger age (254±24 Ma) reflects growth of
793 secondary monazite via dissolution-reprecipitation process. Similar Permian
794 K-Ar ages between 256-268 Ma (K-Ar isochron age, 4 points: 261 Ma) were
795 reported by Depciuch (1972) for KZS diorites from Małolno affected by
796 chloritization/sericitization and interpreted as time of hydrothermal
797 solutions activity. Unaltered samples representing main varieties of KZS
798 igneous rocks provided significantly older K-Ar ages ranging from 291 to
799 303 Ma (K-Ar isochron age, 14 points: 295 Ma).

800 Our mean value of the monazite age (314 Ma) for Szklarska Poręba Huta
801 aplogranite is consistent with high precision ID-TIMS U-Pb zircon ages
802 312.5 ± 0.3 Ma and 312.2 ± 0.3 Ma indicating crystallization of two main
803 granite varieties (porphyritic and equigranular) of the Karkonosze Pluton
804 (Kryza et al. 2014). Similar $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 312 ± 4 Ma reported by
805 Marheine et al. (2002) suggest fast cooling below closure temperature of
806 the K-Ar system. It has to be mentioned, even if of poor quality, an age
807 of 271 ± 20 Ma obtained on monazite from the same aplogranite at Szklarska
808 Poręba Huta (Mikulski et al. 2004). This dating deserves to be linked to
809 the 254 ± 24 Ma obtained on the secondary monazite of the Chwalisław valley
810 section. The Karkonosze granite underwent fast uplift to the surface as
811 evidenced by granitic detrital material in Upper Carboniferous sediments of
812 the North Sudetic Basin (Milewicz 1968, Milewicz 1970).

813 Previous paleomagnetic dating aimed to determine the primary age of
814 the emplacement of several Carboniferous plutonic units of the Polish
815 Sudetes and led to the detection of Permo-Triassic (re)magnetizations (Edel
816 et al. 1997). Even if the study did not plan to determine the age of the
817 albitization, and even if sampling of the obviously altered (reddened)
818 facies was avoided, re-magnetization events, younger than the age of the
819 emplacement of the granitoid, were present in some of the samples from the
820 Chwalisław valley amphibolites in contact with the Kłodzko - Złoty Stok
821 intrusion and in the Iżera dolerite close to the Szklarska Poręba Huta
822 quarry (Edel et al. 1997). The authors identified paleomagnetic carriers of
823 these Permo-Triassic paleopoles as hematite and maghemite, while the

824 Carboniferous paleopoles reported by Edel et al. (1997) are carried by
825 (primary) magnetite. Further paleomagnetic studies on the Kłodzko
826 Metamorphic Complex report on Triassic and Jurassic directions for similar
827 magnetic mineral assemblages (Kadziąłko-Hofmokr et al. 2003).

828 The Permian-Triassic age in the Kłodzko - Złoty Stok massif and the
829 Szklarska Poręba area recurs systematically despite the different dating
830 techniques. Thus, there is consistency between the results of paleomagnetic
831 dating of the reddened/albitized facies in this study and other published
832 age determinations on the various facies which were simply described as
833 'altered' but not clearly characterized as albitized.

834 **Link to the Post-Variscan paleosurface**

835 The Sudetic area, together with the southern part of the Germanic
836 Basin of central Europe, is asssed to have been emergent during Permian and
837 Trias on the basis of stratigraphic and sedimentological evidence (Feist-
838 Burkhardt et al. 2008; Głuszyński and Aleksandrowski 2022). At the same
839 time, Permian and Triassic ZHe cooling ages reported by Danišik et al.
840 (2010) and Sobczyk et al. (2015) are consistent with the idea of exposure
841 of some parts of the Karkonosze granite during the Triassic (see also Migoń
842 and Danišik 2012). Similarly, Carboniferous to Triassic ZHe data reported
843 by Danišik et al. (2012) show that certain parts of the Kłodzko - Złoty
844 Stok granitoid massif may have been close to the Earth surface during the
845 Triassic (249.8 ± 20.0 - 209.4 ± 4.1 Ma) and reflect the record of
846 progressive cooling through the zircon helium partial retention zone.

847 Results of zircon (U-Th)/He (ZHe) dating available for the Western
848 Sudetes area revealed i.a. Triassic (248.2 ± 16.6 - 212 ± 14.2 Ma) cooling
849 signals (Sobczyk et al. 2015). They have been interpreted as a record of
850 bedrock slow exhumation or alternatively as an apparent ages due to zircon
851 thermal system rejunevation, as a result of burial (see discussion in
852 Sobczyk et al. 2020) or an advective heat flow under crustal extensional
853 conditions (Botor et al. 2017). Noteworthy, ZHe datings for the Permian
854 sandstones from the Fore-Sudetic Homocline (Sobczyk et al. 2013) provided a
855 single Mid-Triassic cooling age (244 ± 5.0 Ma) within a Weissliegende
856 series, correlating with paleomagnetic datation which was interpreted, due

857 to its clear post-depositional origin, as a probable age of Kupferschiefer
858 mineralization. Therefore, low-temperature thermochronological data, even
859 if not providing indisputable proof for the Post-Variscan paleosurface
860 exposure during Trias, give some arguments for linking Triassic bedrock
861 cooling with a regional crustal instability during the post-orogenic
862 unroofing. Also the local sedimentological record for the Early Triassic
863 lithofacies within the Intra-Sudetic and North Sudetic synclinoriums
864 indicates channel flow and river bed depositional environments (Kowalski
865 2020), suggesting the predominance of continental conditions for this
866 period with local source areas located mainly in the Karkonosze-Izera
867 Massif. Noteworthy, strongly kaolonized undisturbed Triassic
868 (Buntsandstein) sandstones from the Sudetes are interpreted as a record of
869 *in situ* chemical weathering profiles at the Triassic paleosurface (Kowalski
870 2020 and references therein). Finally, regional review studies indicate
871 that during the Late Triassic to Early Cretaceous the Sudetes landforms
872 were steady sustained by the Post-Variscan paleosurface that formed a vast
873 land mass undergoing slow erosion and denudation resulting i.a. in formation
874 of deep weathering mantles, known also from central and northern Europe
875 (Migoń and Lidmar-Bergström 2001). Powolny et al. (2022) advocated a late
876 to post-magmatic low-temperature fluid-rock interactions controlling the
877 process of albitization within the Permian trachyandesites from the Intra-
878 Sudetic Basin. Based on apatite fission-track central ages obtained for
879 three samples, these authors proposed a Jurassic (182 - 161 Ma) time
880 interval for this thermal event. Closer inspection of apatites analyzed by
881 Powolny et al. (2022) show that >50% of all dated grains revealed a
882 Triassic cooling age, which indicates a possible two stage post-Variscan
883 albitization in the Sudetes.

884 In the absence of any major Triassic magmatic event in the Sudetes
885 and adjacent areas, we link the reddening alteration of the Kłodzko - Złoty
886 Stok and of the Karkonosze granites, as well as their albitization, to a
887 Post-Variscan/Pre-Triassic paleosurface, similar as described for the
888 French Massif Central (e.g. Schmitt and Clement 1989; Parcerisa et al.
889 2009) and in NE Spain (Fàbrega et al. 2019). It follows then that the most
890 pervasively reddened/albitized facies were formed near the paleosurface

891 while at depth the granites were progressively less affected by the
892 alteration process. It is of note that the conspicuous reddened and
893 albitized fractures are never cut by any fractures that do not exhibit this
894 alteration. Hence, we regard the lack of non-albitized fractures as
895 evidence that the alteration is relatively recent and post-dates the
896 exhumation of the granite bodies.

897 Moreover, the superficial origin of albitisation and reddening via
898 weathering explains the spread of the paleomagnetic data along the APWP
899 curve for 100 Ma from the Permian to the Jurassic. This distribution of
900 palaeomagnetic ages is not the result of successive phases of alteration,
901 but rather a single event that is heterogeneously recorded in the rock. It
902 relates to the progression of weathering along fractures and even around
903 kilometer sized massive rock volumes. But presumably there were times of
904 more active and less active phases during this long period, mainly under
905 the influence of climate and eustatic sea level changes, which could
906 explain some clustering of paleomagnetic ages during distinct periods.

907 **INTERPRETATION OF THE PARAGENESES**

908 In general, our observations agree with earlier data on fracture
909 related hydrothermal alterations with the development of albitization and
910 saussuritization (e.g. De Jong and Williams 1995; Lee and Parsons 1997;
911 Engvik et al. 2008; Plümer and Putnis 2009; Sandström et al. 2010; Morad
912 et al. 2010; Powolny et al. 2022). The direct relationship of albitization
913 and related alterations with the Post-Varisacan paleosurface completely
914 changes the paleoenvironmental conditions that prevailed for these
915 alterations compared to quoted studies that ascribe them a deep and
916 relatively high temperature environment. Nevertheless, in the context of a
917 direct relationship with the Post-Varisacan paleosurface, the alterations
918 discussed in this study do not exhibit any typical regolith features, such
919 as illuviation structures related to water percolation in vadose
920 environments, bioturbation, or pallid or iron stained mottles. On the
921 contrary, the isotropic structures around blocks evoke an aquifer
922 environment and recall saprolite from the lower zones of regolith horizons
923 corresponding to deep weathering of bedrock. The most astonishing

924 characteristic is the alteration around the fracture blocks. This occurs
925 without modification of the primary fabric of the rock and without any
926 obvious development of porosity. However, the alteration developed
927 systematically via a hierarchical net of fracturing from the macro-scale,
928 demarking fracture blocks, to intergrain microcracks, giving to the
929 albitization front at outcrop scale a dendritic aspect of fresh unaltered
930 blocks surrounded by reddened fractures (Fig. 9a). Here we examine the
931 geochemical grades, coupled dissolutions, precipitations and water-rock
932 interactions in order to identify the nature of the fluids that triggered
933 these alterations and ultimately the nature of the related
934 paleoenvironment.

935 **Fracture-bound geochemical grades**

936 The alteration primarily and intensively affects the plagioclase and
937 only to a lesser extent K-feldspar and ferromagnesian minerals (biotite and
938 amphibole). The alteration of plagioclase is of three kinds (Fig. 9a and
939 16b).

940 1) Albitization of the primary feldspars is causally linked to proximity to
941 the fractures. Some late Ca-bearing silicates may be associated with
942 this alteration, but they rather fill in microcracks and their origin
943 may not be directly related to the albitized plagioclases.

944 2) Saussuritization of the primary plagioclases accompanied by the
945 development of Ca-bearing prehnite is typical of the non reddened and
946 lightly reddened facies away from fractures. Secondary albite has
947 developed along with saussuritization of the primary plagioclase.

948 3) Invasive development of sericite (sericitization) affects in places the
949 primary plagioclases from the non reddened and lightly reddened facies
950 and is associated with the formation of small neogenic albite but
951 without any Ca-bearing silicates.

952 The alteration of ferromagnesian minerals is connected to the
953 alteration of plagioclase. There is a progressive chloritization of biotite
954 in the non reddened and lightly reddened facies as the fractures are
955 approached. Chloritization seems to occur either under volume conservation
956 or with a slight decrease of the volume of the biotite with secondary

957 minerals such as quartz, K-feldspar, prehnite, and iron-bearing minerals
958 embedded in the cleavage planes. The amphiboles appear to be less sensitive
959 to chloritization than biotite, remaining practically unaltered in the
960 unaltered and lightly reddened saussuritized facies.

961 These alterations result in a strong geochemical overprint between
962 the core and the boundaries of the fracture-isolated blocks (Fig. 9c). The
963 margins of the blocks are strongly Na-enriched due to albitization of
964 plagioclase, and depleted in K due to chloritization of biotite and partial
965 albitization of K-feldspar. The cores of the blocks are relatively enriched
966 in K, Ca and Al compared to the albitized facies (taking into consideration
967 the saussuritization of the plagioclases).

968 The secondary mineral parageneses, especially albitization and
969 saussuritization, never exhibit an interruption or clear alteration front
970 but appear rather as continuum, grading from one to another. There is no
971 obvious temporal succession between the parageneses. By contrast, the
972 parageneses are overlapping, with small sericites in the
973 albitized/hematitized plagioclase crystal cores formed due to alteration of
974 the plagioclase before albitization. The secondary parageneses appear more
975 like an expression of the geochemical gradients between fluids circulating
976 within the fractures and the massive granitoid blocks. Thus, the observed
977 sequences of neogenic minerals result from the progressive encroachment of
978 the reaction front towards the cores of the fracture blocks while their
979 outer margins are gradually altered in equilibrium with the interstitial
980 fluids. It is appropriate at this juncture to refer to metasomatic sequence
981 or zonation rather than paragenesis involving successive or different
982 alteration processes.

983 The albitization as well as the iron oxide pigmentation becomes less
984 intense away from the walls of the fractures towards the inner parts of the
985 fracture blocks. This gradient reflects a decreasing Na⁺ and O₂ (and/or Fe)
986 content of the fluids circulating in the intergranular crystal boundaries
987 (Fig. 9d). In contrast, the K⁺ and Ca⁺⁺ content of the interstitial fluids
988 would have decreased in the opposite direction: K-bearing minerals (K-
989 feldspar, biotites) and Ca-bearing plagioclases were altered by leaching

990 of K⁺ and Ca⁺⁺ at the fracture walls and sericite formed within the lightly
991 reddened and unaltered facies progressively more distant from them.

992 **Coupled dissolution-precipitation reactions**

993 It is well known, plagioclase with a high An content is susceptible
994 to preferential dissolution due to enhanced instability in the presence of
995 water (Matthews and Goldsmith 1984; Stillings and Brandtley 1995). Thus,
996 destabilization of plagioclase could be a trigger resulting in a chain of
997 reactions in the altered granitoids.

998 Albitization of plagioclase proceeds at constant volume (with low
999 porosity) given a Na-supply, additional silica, and with the release of Ca
1000 and some Al (Ramseyer et al. 1992; Parcerisa et al. 2009; Hövelmann et al.
1001 2010). In the vicinity of the fractures, the released cations would have
1002 been leached and evacuated by the circulating fluids. As the reaction
1003 continued, fluid circulating in the rock by diffusion and convection would
1004 have become gradually depleted in Na, and albitization diminished and or
1005 even ceased at low Na-content (Fig. 9d).

1006 Biotites are destabilized and transformed into chlorite in the
1007 vicinity of the fractures, probably through a double effect of the leaching
1008 of the K interlayer cations and oxidation of the Fe²⁺ in octahedral sheets
1009 (Fig. 9e). Here leaching of Si(OH)₄ takes place because the cation/Si ratio
1010 is equal to ~1 within the biotite, whereas it is between 1 and 1,5 within
1011 the replacement chlorite. Detailed study of the Karkonosze granite's
1012 chloritization (Wilamowski 2002) showed that "The rate of potassium removal
1013 from biotite during the progress of chloritization is not constant. The
1014 rapid removal of K is observed at initial stages of chloritization. After
1015 the replacement of 20% of the biotite, K removal is much slower." The
1016 released silica could have nourished the albitization of the plagioclase.
1017 If the chloritization occurred without an additional cation supply, the
1018 volume loss would have been about 30%, whereas if additional cations were
1019 introduced (such as the Al released by plagioclase alteration) the volume
1020 loss may have been less substantial (Ferry 1979).

1021 Amphiboles appear to be less "sensitive" to chloritization than the
1022 biotites, probably because their structure does not contain any unstable or

1023 mobile cations such as K^+ . The destabilization of the amphiboles is most
1024 likely a result of leaching of Mg and perhaps oxidation of Fe^{2+} . It might be
1025 that amphiboles, and in particular hornblendes (which are calcic), are
1026 relatively protected against chloritization because of the Ca liberated by
1027 albitization of the plagioclase. For that reason the chloritization of
1028 amphiboles occurs only in the more intensively altered facies in which
1029 plagioclases are albitized and therefore no longer provide Ca^{++} to the
1030 fluids.

1031 Hematitization is systematically associated with the albitization.
1032 Hematite occurs mainly as pigments within the neogenic albite or as coarser
1033 crystalline material associated with the ferromagnesian minerals. There is
1034 no iron oxide precipitation in the fractures, so the fluids circulating in
1035 the fractures appear unlikely to have been iron-bearing. Hematite has
1036 likely been nourished by the iron released by alteration of ferromagnesian
1037 minerals.

1038 The formation of sericite depended mainly on the availability of K^+ .
1039 Chloritization of biotite was the quasi-exclusive source of K in the system
1040 because albitization of K-feldspar was not pervasive. Besides, K released
1041 by chloritization in the early stage of the alteration and in the vicinity
1042 of fractures would have been evacuated by convective fluid flow.
1043 Sericitization is present away from the fractures where diffusion was
1044 dominant with respect to the movement of the chemical elements. Thus,
1045 sericitization is dominant in the non reddened and moderately reddened
1046 facies where the Na-content was lowered in the interstitial fluids due the
1047 increasing distance from the fractures that are the major source of Na.
1048 Yet, where the K/Na ratio was higher, albitization was limited and K
1049 competed with Na to form K-bearing silicates (mainly secondary K-feldspars
1050 and/or sericite) in combination with Si and Al (Ben Baccar et al. 1993;
1051 Perez and Boles 2005).

1052 Euhedral crystals of prehnite within altered plagioclase indicate
1053 epigenetic replacement with the main chemical component directly supplied
1054 in situ by the plagioclase. On the other hand, prehnite fining in
1055 microfractures witnesses the mobility of these components over some

1056 distance. If the mobility of Ca and Si is to be expected, it is rather
1057 unusual for Al , typical fairly immobile in a medium close to neutral pH,
1058 except in presence of complexing agents (Castet et al. 1993; Farmer and
1059 Lumsdon 1994 ; Oelkers et al. 1994). In any case, Al appears to have been
1060 fairly mobile, at least in some cases, but it is difficult to assess
1061 whether or not Al was evacuated from the fractures by fluid circulation.

1062 Recrystallization of monazite was also apparently triggered by
1063 calcium availability. Initially, primary monazite recrystallized into
1064 secondary monazite at the early stages of albitization but later, during
1065 advanced stages of the albitization, primary and secondary monazite
1066 recrystallized into apatite in concert with the increase in calcium
1067 availability in the system.

1068 These alterations can be considered to be topochemical: the
1069 saussuritization of plagioclase provided Ca for prehnite development and
1070 the chloritization of biotite is the likely source of K for sericitization.
1071 However, the complete albitization of the plagioclases and in particular
1072 the albitization of the K-feldspars in the vicinity of the fractures
1073 testify for the supply of Na, and the hematite associated with albite
1074 registers the oxidative capacity of the alteration and therefore a supply
1075 of oxygen.

1076 **Fluid-rock interactions**

1077 The geochemical exchanges between reaction zones are determined by
1078 the transport of the ions in the fluids. Exchanges in fractures record the
1079 circulation of the transport fluids by convection. The circulation may have
1080 been relatively rapid in the major fractures but slower as the dimensions
1081 of the fractures decreased. Away from the open micro-fractures, the
1082 quantity of fluid is reduced and mainly limited to 'thin films' at the
1083 boundaries of mineral grains and the dissolution-recrystallization
1084 interface. Such confined fluids may be very different in chemical
1085 composition and behavior from bulk fluids (Putnis and Ruiz-Agudo 2013).
1086 Within thin films, the circulation would have been minimal, diffusion
1087 prevailed over convection and ion mobility was extremely reduced.

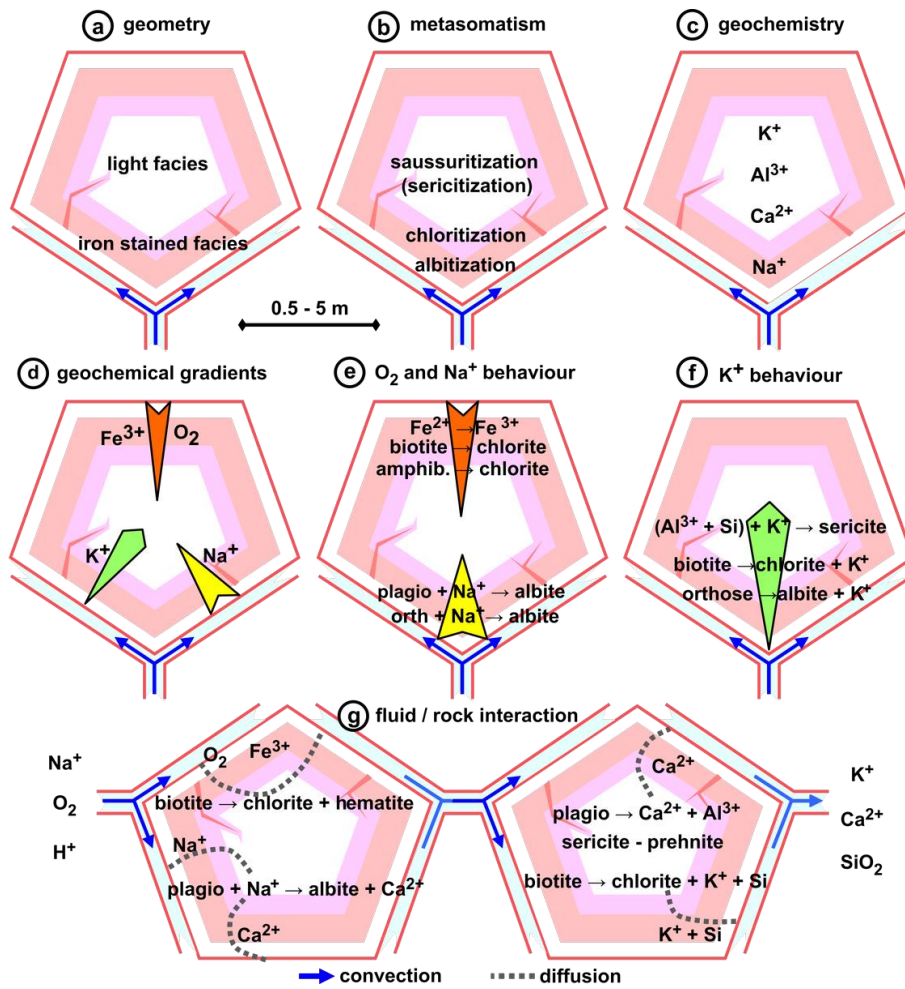
1088 In this context, the coupled-reactions described above are not part
1089 of a global geochemical equilibrium related to the fluids in the fractures
1090 but they are rather part of a spatially-linked dynamic geochemical
1091 equilibrium. The balances are not determined by the kinetics of the
1092 mineralogical reactions but rather by the kinetics of the diffusion of ions
1093 in the rock. The system was only partially open. A geochemical zoning
1094 developed and progressed in a chromatographic way: with displacement of
1095 ions at diverse rates and according to fronts governed by distinct
1096 reactions. This is characteristic of a metasomatic system (Korzhinskii and
1097 Oestreich 1965; Brinhall and Dietrich 1987; Guy 1993). Thus, thermodynamic
1098 reactions that originally seem incompatible with a global context may
1099 occur. Temperature is not a limiting factor but it allows speeding up of
1100 the reactions by accelerating diffusion. The reaction rate is controlled by
1101 the diffusion rate and is extremely slow, but not impossible from a
1102 thermodynamical point of view.

1103 Thus, within a block of massive rock bounded by fractures, a zoning
1104 of the ion transport that determines the geochemical zoning is primary
1105 established (Fig. 9g). Along the fracture walls, exchanges by convection
1106 prevail and are relatively fast; the mineralogical parageneses formed in
1107 equilibrium with the fluid and thereby reveal the geochemical
1108 characteristics of the fluid. Further away from the fracture walls,
1109 exchanges are slow and are driven by diffusion; the mineralogical
1110 parageneses are controlled by local balances that are far from the balance
1111 within the fluid. This alteration of massive blocks differs entirely from
1112 alteration of porous rock where ion mobility is practically exclusively
1113 provided by fluid flowthrough.

1114 **Geochemistry of the fracture fluids.**

1115 Albitization of plagioclase and K-feldspar reflects an interaction
1116 with Na-enriched and relatively K-depleted fluids (Saigal et al. 1988;
1117 Aagaard et al. 1990; Ben Baccar et al. 1993; Schmitt 1994; Perez and Boles
1118 2005). The association with hematite indicates that the fluids were also
1119 enriched in oxygen. There seems to have been a relatively low CO₂ fugacity
1120 since only minor calcite was formed, in spite of the abundance of Ca

1121 released during plagioclase alteration. The fluids were also relatively
1122 under-saturated in SiO₂, Al and Mg, since chlorite coating the fractures is
1123 sometimes entirely altered to hematite. On exiting the system, fluids were
1124 depleted in Na and O₂. However, they were enriched in K because, in the
1125 early stages of the alteration, little or no sericite was formed within
1126 plagioclases during the chloritization of biotite. The chemical balance is
1127 more difficult to establish for SiO₂, Al and Ca. Significant amounts of
1128 silica were liberated during chloritization of biotite so it is likely that
1129 it was exported from the system, at least in the granodioritic facies that
1130 is rich in biotite. It could be that Al and Ca were in balance between the
1131 altered plagioclases and the formation of prehnite and calcite, even if a
1132 quantified balance is not available. Nevertheless, it is obvious that the
1133 Ca was easily exported from the cortical reactional zones during the early
1134 stage of the alteration. The mobility of Al is unlikely to have occurred on
1135 a significant scale because the major fractures never show any aluminosilicate
1136 mineralization. In conclusion, it appears clear that the fluids
1137 responsible for the alteration were oxidized, Na-rich and relatively
1138 depleted in K in comparison to Na.



1139

1140 **Fig. 9** Schematic interpretative sketch of the various petrographic
 1141 parageneses related to reddening and albitization of the granitoid rocks in
 1142 the Sudetes at the outcrop scale

1143

1144 **Iron oxide behavior**

1145 Maghemite is considered to be metastable with respect to hematite
 1146 (Waychunas 1991; Majzlan et al. 2003). In addition, it is generally
 1147 admitted that slow and incomplete oxidation of magnetite favours the
 1148 topotactic formation of maghemite whereas faster oxidation of magnetite,
 1149 under highly oxidizing conditions, results in hematite formation
 1150 (Steinthorsson and Helgason 1992; Bowles et al. 2011). The maghemite-
 1151 hematite transition during albitization may be attributed to changes in the
 1152 alteration environment during the oxidation of magnetite. In this case the
 1153 alteration commenced with the formation of maghemite under low oxidizing
 1154 conditions in the non-reddened and slightly reddened granitoids and

1155 continued with the formation of hematite in the reddened facies under
1156 stronger oxidizing conditions, when ferrous silicates (biotite and
1157 amphibole) were being oxidized. Thus, the occurrence of maghemite and
1158 hematite suggests that the oxidizing conditions changed during the
1159 alteration of the granitoid and the spatial distribution of hematite and
1160 maghemite provides a tool to estimate the gradient of these conditions. The
1161 distribution of the alteration environments substantiates the idea that
1162 reddened albitized facies and non reddened saussuritized/sericitized facies
1163 reflect gradational changes during a single event.

1164 From a geochemical point of view there is no contradiction in the
1165 spatial discrepancy between the zones of alteration of the silicate
1166 minerals and the distribution of the iron oxides. The physico-chemical
1167 properties that constrain their reactions are not the same. A key-factor in
1168 the alteration and crystallization of the silicates is the pH, whereas Eh
1169 and the oxygen fugacity controls the oxidation state of the iron oxides.
1170 These two parameters may vary independently. Nevertheless, the
1171 precipitation of iron oxides is also facilitated by the elevated pH. Thus,
1172 local lowering of the H⁺ activity at the sites of feldspar alteration will
1173 favour the localized precipitation of hematite as inclusions within the
1174 feldspars (Boone 1969). This also explains the local distribution of
1175 pigmentary hematite inclusions in albite during the alteration of both
1176 plagioclase and K-feldspar. On the other hand, the intensity of the
1177 reddening, or the abundance of the pigmenting hematite, is related to
1178 oxygen fugacity and therefore is likely to decrease away from the
1179 fractures. From the geochemical point of view, the hematite distribution is
1180 consistent with the alteration sequence of the silicate minerals.

1181 **Origin of the oxidizing fluids**

1182 In general, post-magmatic alterations are explained by a hydrothermal
1183 model which uses large convective circuits to bring the superficial fluids
1184 in contact with the warm rock complexes, in particular granitic intrusions,
1185 in which the alterations then take place (Taylor 1977; Jenkin et al. 1992;
1186 Putnis et al. 2007; Drake et al. 2008). The magmatic intrusions provide
1187 heat to the fluids and the subsequent loss in fluid density causes the

1188 upwelling, maintaining the circulation between the landsurface and depth.
1189 Fluids with a deeper origin may be also mixed into this circulation model.

1190 In the hydrothermal model, the superficial origin of the fluids, or
1191 at least the participation of meteoric water, has been evoked by oxygen and
1192 hydrogen isotope studies on the alteration minerals (Simon and Hoefs 1987;
1193 Sun and Eadington 1987; Jenkin et al. 1992; Boyce et al. 2003). These
1194 models usually consider superficial oxygen-rich fluids descending very
1195 deeply (more than 10 km) before they interact with the rock matrix.
1196 However, some reactive transport models strongly suggest that this form of
1197 fluid percolation leads to a rapid depletion of the oxygen fugacity in the
1198 circulating fluids, so fluids at great depths are very likely to be under-
1199 saturated in oxygen (MacQuarrie et al. 2010).

1200 Here we have documented the Permian-Triassic-Jurassic age of the
1201 alteration and provided arguments that it is necessarily related to the
1202 Post-Variscan paleosurface and *de facto* involved infiltration of weathering
1203 solutions and/or groundwater flow.

1204 **Temperature and depth of formation**

1205 The $\delta^{18}\text{O}$ results allow to estimate the fluid temperature during
1206 plagioclase albitization and K-feldspar microclinization. Assuming a local
1207 isotopic equilibrium between minerals and the fluid at the reaction
1208 interface, the isotopic mass balance was calculated for each reaction using
1209 the equation of Taylor (1977), by (Eq. 1):

$$\delta^{18}\text{O}_f^w = \delta^{18}\text{O}_i^w + \frac{R}{W} \times (\delta^{18}\text{O}_i^m - \delta^{18}\text{O}_f^m) \quad (1)$$

1210 Where $\delta^{18}\text{O}_i^w$ and $\delta^{18}\text{O}_f^w$ are the initial and final oxygen isotope
1211 compositions of the fluid, respectively, R and W the percentage of oxygen
1212 atoms in the rock (or mineral) and water, respectively, and $\delta^{18}\text{O}_i^m$ and $\delta^{18}\text{O}_f^m$
1213 the oxygen isotope composition of the initial and final minerals,
1214 respectively. For the rock-to-water oxygen mass ratio (R/W), conservative
1215 estimates of 0.5 and 1 were applied. The applied values for the oxygen
1216 isotope composition of the initial fluid ($\delta^{18}\text{O}_i^w$) were -8 and -12‰ (VSMOW),
1217 which are typical $\delta^{18}\text{O}$ values observed in deep groundwaters hosted in stable

1218 granitic basements (Stober & Bucher 1999a and b; Barth 2000; Kloppmann et
1219 al. 2002).

1220 Two options of reactions paths were considered, (1) sequential
1221 reactions, with a fluid moving across a steeped sequence of albitization
1222 and microclinization, so that the mass balance of albitization is
1223 calculated first and subsequently the mass balance of microclinization is
1224 calculated using as initial fluid isotopic composition the $\delta^{18}\text{O}$ value of the
1225 final fluid obtained from the mass balance of previous albitization or, (2)
1226 concomitant reactions, were the fluid is renewed at the beginning of each
1227 reaction. After the isotopic mass balance estimation for each reaction, a
1228 temperature was calculated for each reaction using the plagioclase-water
1229 (Eq. 2) and K-feldspar-water (Eq. 3) equilibrium fractionation equations of
1230 Zheng (1993).

$$1000 \ln \alpha_{Ab-H_2O} = \frac{4.33}{T^2} \times 10^6 - \frac{6.15}{T} \times 10^3 + 1.98 \quad (2)$$

$$1000 \ln \alpha_{Kfs-H_2O} = \frac{4.32}{T^2} \times 10^6 - \frac{6.27}{T} \times 10^3 + 2.00 \quad (3)$$

1231 Using the two reaction paths and the crossed combinations of the
1232 different applied values for R/W and $\delta^{18}\text{O}^w_i$, the estimated interaction
1233 temperatures of water during albitization and microclinization range
1234 between 32 and 89°C (average of 60°C). Significantly, and despite the large
1235 relative variations in the values applied to R/W and $\delta^{18}\text{O}^w_i$, and the two
1236 different reaction paths, the temperature estimates present a narrow range
1237 of variability, supporting that the mineral reactions developed in by
1238 interaction with "low" temperature groundwaters. Thus, considering the
1239 Permian-Triassic age and the relatively low temperature of the alteration,
1240 the hypothesis that the alteration fluids were closely related to the
1241 paleosurface and *de facto* "vertical" infiltrations of a weathering solution
1242 is a plausible possibility. Similar reaction temperatures (45 to 65°C) were
1243 obtained for plagioclase albitization and K-feldspar microclinization in
1244 Permian-Triassic age red-stained albitized profiles in NE Spain (Fàbrega et
1245 al. 2019), that are similar in field layout, petrographic characteristics
1246 and age to the Polish sections of this study.

1247 Regardless of the lack of accuracy of these temperature estimates,
1248 these estimated values show that the temperatures of the reactions
1249 associated with the development of red-stained albitized rocks are
1250 significantly below temperatures suggested for albitization formed under
1251 hydrothermal alteration (Cathelineau 1986; Boulvais et al. 2007) and tardi-
1252 magmatic alteration (Lee and Parsons 1997; Fiebig and Hoefs 2002). Instead,
1253 the low temperature range accords with the geological settings, age and
1254 association of the red-stained albitized profile developed on the Post-
1255 Variscan paleosurface during the Permian-Jurassic period. We suggest that
1256 the albitized profile developed a few hundred metres beneath the Post-
1257 Variscan paleosurface by interaction with relatively 'low' temperature
1258 groundwaters, and was subsequently exposed at the landsurface after erosion
1259 of the uppermost weathering mantle.

1260 **PALEOENVIRONMENTAL SETTING OF THE ALBITIZATION**

1261 The absence of mesoscale- and microscale-geopetal structures,
1262 together with clues of directional water flow combined with the isotropic
1263 structures around fracture blocks within the macro-scale profiles, evoke an
1264 aquifer environment at the origin of the albitization. Boreholes in the
1265 Massif Central (Schmitt 1986) and field exposures in the Morvan (Parcerisa
1266 et al. 2009) and NE Spain (Fàbrega et al. 2019) show that the albitized
1267 profiles may generally reach 100 m in thickness and may even exceed 200 m.
1268 Such profiles are comparable to those of weathering profiles with thick
1269 basal saprolite horizons in tropical soils (Boulangé 1984; Vasconcelos et
1270 al. 1994; Ruffet et al. 1996; Grimaud et al. 2015) and deep groundwater
1271 saprolites in dry saline environments as well as in periglacial
1272 environments (Thiry et al. 2006; Thiry et al. 2017). Nevertheless, the
1273 primary profiles were probably much thicker because only the deep indurated
1274 horizons have survived the erosion that preceded the Mesozoic
1275 transgression. The profiles were certainly at some stage surmounted by non-
1276 indurated horizons and regolith materials. On the other hand, the
1277 functioning of the albitization profiles and thus landscape stability
1278 spreading over ~50 Ma in the Sudetes is comparable to the duration of the

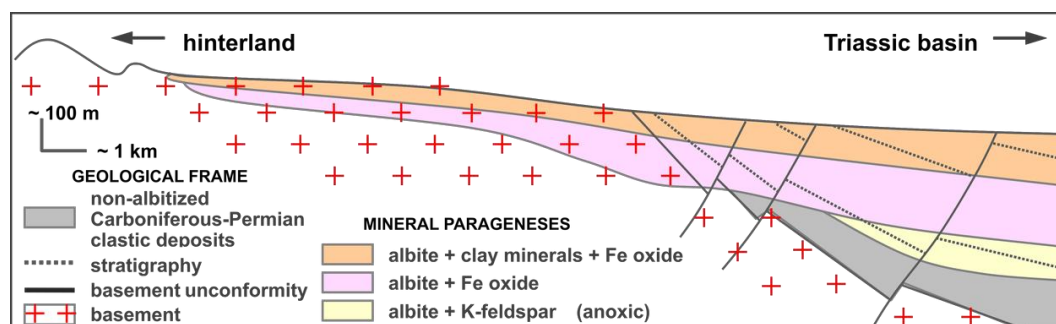
1279 deep tropical weathering profiles and related landscapes rooted in old
1280 cratons mentioned above.

1281 Reaction temperatures calculated from the oxygen isotope composition
1282 of minerals, and hypotheses about the oxygen isotope composition of
1283 reactant fluids and their circulation, led us to an estimation of the mean
1284 temperature of albitization and associated reactions of about 60°C in the
1285 Sudetes albitized profiles and 55°C in the Catalan albitized profiles
1286 (Fàbrega et al. 2019). Based on this temperature, and considering a
1287 geothermal gradient of about 25–30°C km⁻¹ and a surface temperature of
1288 ~30°C, in consideration of the high paleotemperature proposed at the
1289 Permian-Triassic boundary, led to calculate a mean depth of about 1 km for
1290 the environment in which albitization occurred.

1291 There is indeed a systematic vertical profile organization for the
1292 sections we have studied, specifically with a more albitized and
1293 ferruginized upper part, a less-altered and reddened lower part where
1294 albitization occurred under weakly oxidizing conditions, and the
1295 progressive weakening and disappearance of alteration with depth. In
1296 connecting this geometry to the Pre-Triassic paleosurface, it is clear
1297 that the albitization represents a climatic weathering process of great
1298 magnitude. The low mobility of most of the involved chemical elements could
1299 point to deep groundwater table with very low outflow. But even if the
1300 alteration occurred at a greater depth, reactant solutions would have to
1301 have been renewed to provide at least sufficient O₂ and Na⁺. Therefore,
1302 there must have been water flow within fractures between the fractured
1303 blocks which implies a hydraulic gradient. In particular, a regional
1304 hydraulic gradient with meteoric water infiltrating the upstream profiles
1305 and groundwater outflowing downstream would have been necessary. As
1306 albitization developed, water changed composition from entrance to outflow
1307 with loss of O₂, Na⁺, and H⁺ and a likely increase in K⁺, Ca⁺⁺, and possibly
1308 SiO₂ content, as deduced from the geochemistry of the mineral
1309 transformations (Fig. 9g). The lateral evolution of such albitized profiles
1310 has been documented in the SW of the French Massif Central (Yerle and Thiry
1311 1979; Clément 1986; Schmitt 1986). Here, (1) profiles on basement heights
1312 are thin and albitization is mainly restricted to fracture walls (such as

1313 described herein), and (2) lowland profiles rooted in Carboniferous and
 1314 Permian deposits are much thicker with upper horizons albitized and
 1315 oxidized whereas deeper albitized horizons remain anoxic and are the site
 1316 of K-feldspar neof ormations (Fig. 10).

1317



1318

1319 **Fig. 10** Arrangement of mineral parageneses on the southern edge of the
 1320 Massif Central. Albitized profiles thicken from crystalline heights towards
 1321 Carboniferous and Permian basin lowlands where K-feldspar forms at depth.
 1322 Modified after Clément (1986)

1323 Two fundamental elements must be taken into account: (1) the presence
 1324 of relief to ensure the hydrological gradient, and (2) the origin of sodium
 1325 that triggers the whole chain of alteration. In terms of the groundwater
 1326 dynamics, it could be that albitization developed in landscapes of
 1327 intermediate relief, perhaps on large piedmonts between a mountain range
 1328 and sea level. Petrographic data indicate that sodium was imported without
 1329 potassium enrichment. Early modeling attempts (Schmitt 1994) also indicated
 1330 that a high Na^+/K^+ ratio was probably required. The Na^+ enrichment is most
 1331 likely to be linked with the peculiar geochemical setting of the Permian-
 1332 Triassic environments in the region. Sodium could have originated from the
 1333 gigantic salt deposits of these periods and may have been imported in large
 1334 quantities and over long distances by successive phases of eolian reworking
 1335 from littoral and endoreric basins during dry periods (Bourquin et al.
 1336 2011).

1337 It should be stressed that the albitization started long before the
 1338 recorded dating, possibly as early as Mid-Permian, with the development of
 1339 the Zechstein evaporite deposits in the Germanic basin. It is the final

1340 stage that we have dated: previous stages have likely been overprinted by
1341 succeeding reactions. Classically, weathering profiles progress from top to
1342 base and result in older ages in the uppermost horizons and younger ages at
1343 depth, as evidenced in numerous long-lasting paleoweathering sequences
1344 (Vasconcelos et al. 1994; Dammer 1995; Ruffet et al. 1996). However,
1345 paleomagnetic dating of the profiles we have studied here show an inverse
1346 polarity, with the oldest ages in the weakly albitized and reddened bottom
1347 horizons and the youngest ages in the more albitized and ferruginized
1348 uppermost horizons. This points to a 'regressive' weathering sequence, in
1349 which the weathering profile did not continue to deepen but was 'blocked'
1350 with only the upper horizons continuing to be weathered. This profile form
1351 relates in some way to the specific evolution of the landscape in the
1352 region in relation to the widespread Mesozoic transgression. It would have
1353 corresponded to a rise in basal hydrological level which induced rising
1354 watertables across the continent. The rising watertables would have limited
1355 the infiltration of surficial waters, changed the hydrodynamic gradients
1356 and effectively stopped the alteration at depth. Thus, the regional
1357 geodynamics during the Late Permian and Triassic period strongly influenced
1358 weathering and regolith processes while the climate was steadily warm and
1359 semi-arid.

1360 Weathering toposequences should be considered on a continental scale
1361 as shown by Fàbrega et al. (2019). In the case of our study area, the
1362 toposequence is arranged between the relief of the Variscan Belt that
1363 provided the hydraulic gradient and the Permian and Triassic basins that
1364 probably supplied the salt. Albitization was initiated and probably
1365 gradually invaded the entire Post-Variscan paleosurface across several
1366 10's of Ma during which both landscape and climate were relatively stable.
1367 Indeed, the overall ages determined up to now for the albitization span a
1368 relatively long period, between 270 and 170 Ma, that corresponds to the
1369 Middle Permian to the Middle Jurassic (Table 4). The distribution of the
1370 now-dated albitized profiles along the Post-Variscan paleosurface between
1371 the Zechstein Sea in the North and the Tethys in the South (Fig. 11), shows
1372 that:

- 1373 4) the profiles closest to the marine environments, namely the Sudetes and
1374 NE Spain, show the oldest ages (this study; Fàbrega et al. 2019);
- 1375 5) the Vosges massif, which was connected to the marine environment through
1376 the threshold passage of the Triassic transgression coming from the
1377 North, also show relatively old albitized profiles (Vercruyssen 2011);
- 1378 6) the area most distant from the sea, namely the Rouergue heights overlain
1379 by infra-Jurassic deposits, comprises the youngest albitized profiles
1380 (Schmitt et al. 1984) while the Permian graben covered with coarse
1381 Triassic deposits on lower plateaux in Rouergue (Bonhomme et al. 1980)
1382 and Morvan (Ricordel et al. 2007) have older ages; and
- 1383 7) in the Sudetes, the highly evolved ferruginous albitized facies
1384 currently in hillside positions are relatively young while the valley-
1385 bottom sections are older. This differentiation is possibly related to
1386 local morphologies whereby raised tectonic blocks remained emerged
1387 during the Mesozoic transgression.

1388 The concordance between age and relative position in the Triassic
1389 paleogeography is striking. The areas covered early by the Triassic
1390 transgression have the oldest albitized profiles whereas the highlands
1391 distant from the sea have the youngest ages. It is actually the
1392 interruption of weathering by the Triassic transgression that is dated.
1393 These datings could in future be used as a tool to better enhance the
1394 mapping of the Triassic transgression and the geodynamics of the basement
1395 as well as its post-Triassic evolution.

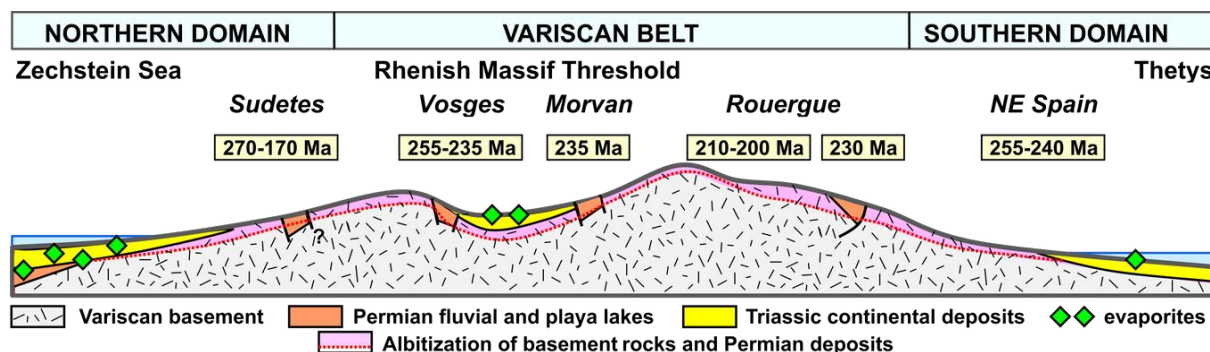
1396 Paleolandscape and paleoweathering features of the Post-Variscan
1397 paleosurface in Western Europe may be compared to regolith features
1398 existing today on the African, Australian and South American cratons
1399 inherited from Pangea break out. Combined radiogenic dating and
1400 geomorphological analyses allowed us to document paleolandscape and
1401 paleoweathering profiles that functioned for more than 100 Ma in these
1402 cratonic regions (Vasconcelos et al. 1994; Dammer 1995; Ruffet et al. 1996;
1403 Pillans, 2007; Beauvais et al. 2008; Peulvast et al. 2008 ; Beauvais and
1404 Chardon 2013; Jean et al. 2020). This is the spatial and temporal scale of
1405 the albitization and hematization paleoweathering profiles we studied on

1406 the Western European Post-Variscan paleosurface. Persistence of the Post-
 1407 Variscan paleosurface is conformed by the widespread magnetic overprinting
 1408 phase due to secondary hematite formation (related or not to albitization)
 1409 that affected widely the Variscan belt during the 235-250 Ma time-span Edel
 1410 et al. (2018).

1411 Table 4 - Age of the reddened/albitized rocks related to the Post-Variscan
 1412 paleosurface on the Variscan Belt

area	location	geomorph	rockl	sample	method	age	reference
Sudetes	Szklarska Poręba	hill	granitoid	2.5 x 2.5 cm	palaeomag	256-190	herein
	Laski quarry	valley	granitoid	2.5 x 2.5 cm	palaeomag	270-190	herein
	Laski village	hill slope	granitoid	2.5 x 2.5 cm	palaeomag	258-170	herein
	Chwalisław	valley	granitoid	2.5 x 2.5 cm	palaeomag	205-195	herein
	Chwalisław	valley	granitoid	monazite	U-Th-Pb _{tot}	254	herein
	Chwalisław	valley	granitoid	2.5 x 2.5 cm	palaeomag	250	Edel et al. (1997)
Vosges			granitoid	2.5 x 2.5 cm	palaeomag	255-235	Vercruysse (2011)
Morvan			granitoid	2.5 x 2.5 cm	palaeomag	235	Ricordel et al. (2007)
Rouergue	Bertholène	high	orthogneiss	minerals	K-Ar	210-200	Schmitt et al. (1984)
	Brousse	basin	sediments	clay fraction	K-Ar	230	Bonhomme et al. (1980)
NE Spain	Guilleries Massif		granite	monazite	U-Th-Pb _{tot}	251	Fàbrega et al. (2019)
	Roc Frausa Mass.		granite	monazite	U-Th-Pb _{tot}	254	Fàbrega et al. (2019)
	Hortmoier Massif		granite	K-F crystals	K-Ar	240	Fàbrega et al. (2019)

1413



1414

1415 **Fig. 11** Position and age of the studied albitized profiles on a schematic
 1416 N-S geomorphological section of the Post-Variscan paleosurface and

1417 associated sedimentary basins during Permian-Triassic times. The oldest
1418 ages occur nearest the sea and youngest away the sea and in highland areas
1419 that were later buried. Paleogeography after Bourquin et al. (2011)

1420 **CONCLUSIONS**

1421 The oxidized and albitized facies of the Polish Sudetes granitoids
1422 exhibit two types of 'classical' petrographical parageneses: a
1423 saussuritization-sericitization and an albitization-hematitization. The
1424 chloritization of amphiboles and biotites that is associated to these
1425 alterations is intermediate between both. These alterations are always
1426 closely associated without any break or spatial discordance between them.
1427 They are controlled by the presence of fractures and are spatially arranged
1428 with respect to the configuration of these fractures. Albitization and
1429 hematitization are preferentially intense close to the walls of the
1430 fractures whereas the saussuritization and the sericitization of
1431 plagioclase become more important further away from the fractures and
1432 towards the center of the massive blocks. These alterations can be
1433 considered as topochemical. However, the complete albitization of the
1434 plagioclase and K-feldspar in the vicinity of the fractures signifies a
1435 supply of Na. Hematite associated with albite is an indication of oxidative
1436 capacity and therefore the supply of oxygen via the alteration.

1437 Paleomagnetic dating of the reddened and albitized facies reveal Late
1438 Permian to Early Jurassic ages for the iron-oxides and radiochronological
1439 dating of associated monazite using the U-Th-Pb_{total} method indicates Late
1440 Permian recrystallization ages. Given the ages of the alteration, it can
1441 clearly be related to the extensive Post-Variscan paleosurface in the
1442 region, since the altered granitoids were already outcropping during from
1443 Late Permian to Early Jurassic. Therefore, these reddened and albitized
1444 profiles can be considered a consequence of deep weathering.

1445 As there was clearly a requirement to renew solutions to feed
1446 chemical reactions, it is likely that the alteration profiles were
1447 downstream of highlands in order to generate a hydrological head. The
1448 sodium supply that triggered albitization could well have come from the

1449 gigantic salt playas that characterized the lowland paleogeography of the
1450 Permian and Triassic in the region. Thus, we suggest that the reddened and
1451 albitized profiles developed on a wide glacis shaped in the basement rocks
1452 upstream of the evaporitic basins. The deep weathering was interrupted and
1453 preserved by the Mesozoic transgression.

1454 The age dating shows that the oldest examples of the alteration
1455 profiles (Late Permian to Early Triassic) are proximal to the Zechstein and
1456 Tethys seas, while the youngest (late Triassic to basal Jurassic) are
1457 further away and appear to be located on the highlands that were more
1458 recently covered by the Mesozoic transgression. Thus, this alteration of
1459 the crystalline rocks is an important marker to constrain Triassic and
1460 post-Triassic geodynamics in the region, including erosion rates and the
1461 geomorphic evolution of the Paleozoic massifs.

1462 Albitized and oxidized facies of crystalline basement are widespread
1463 throughout the world. Until now, this subject has attracted relatively
1464 little interest from petrographers. A re-examination of all existing data
1465 could involve a reconsideration of questions such as prehnite and sericite
1466 paragenesis, particularly from the viewpoint of temperature of formation,
1467 since they have in the past been used as an indicator of 'high
1468 temperatures'. It may be that the formation of sericite and prehnite is not
1469 part of a unique alteration leading to albitization, but the fact remains
1470 that proximity to the paleosurface is clearly indicated by the presence of
1471 hematite.

1472 We suggest that particular attention should be paid to the
1473 radiochronological and paleomagnetic dating results that some have
1474 considered to be "obviously too young" in the context of the age of
1475 emplacement of the crystalline massifs. These "too young" dates have been
1476 systematically rejected as false in the past and rarely noted in
1477 publications; consequently this information was lost. A new consideration
1478 of these dates may realise their consistency, as shown for example by Edel
1479 and Schneider (1995) for the Permian formations that endured a Triassic
1480 rejuvenation highlighted by paleomagnetism.

1481 It should be noted that the formation of secondary apatite in the
1482 reddened albitized Sudetes rocks has previously been reported in
1483 conjunction with albitization and chloritization related to the Triassic
1484 paleosurface in the French Massif Central (Parcerisa et al. 2009) and in NE
1485 Spain (Fàbrega et al. 2019). It also formed during metasomatic albitization
1486 in a metagabbro from Norway (Engvik et al. 2009). The formation of
1487 significant amounts of secondary apatite during albitization must be taken
1488 into account if fission track analysis or (U-Th)/He dating is carried out
1489 on these reddened albitized rocks.

1490

1491 **References:**

1492

1493 Aagaard P, Egeberg PK, Saigal GC, Morad S, Bjørlykke K (1990) Diagenetic
1494 albitization of detrital K-feldspars in Jurassic, Lower Cretaceous,
1495 and Tertiary clastic reservoir rocks from offshore Norway, II.
1496 Formation water chemistry and kinetic considerations. Journal of
1497 Sedimentary Petrology 60:575-581

1498 Aleksandrowski P, Kryza R, Mazur S, Żaba J (1997) Kinematic data on major
1499 Variscan strike-slip faults and shear zones in the Polish Sudetes,
1500 northeast Bohemian massif. Geological Magazine 133: 727-739

1501 Aleksandrowski P, Mazur S (2002) Collage tectonics in the northeasternmost
1502 part of the Variscan Belt: the Sudetes, Bohemian Massif). In:
1503 Winchester, J., Pharaoh T. & Verniers J. (eds), Palaeozoic
1504 Amalgamation of Central Europe, Geological Society, London, Special
1505 Publications 201: 237-277

1506 Alt, J.C., Laverne, C., Coggon, R.M., Teagle, D.A., Banerjee, N.R., Morgan,
1507 S., Smith-Duque, C.E., Harris, M., Galli, L. (2010) Subsurface
1508 structure of a submarine hydrothermal system in ocean crust formed at
1509 the East Pacific Rise, ODP/IODP Site 1256. Geochemistry, Geophysics,
1510 Geosystems, 11(10)

1511 Awdankiewicz M, Awdankiewicz H (2010) Spessartites of the Kłodzko Złoty-
1512 Stok dyke swarm. Mineralogia Special Papers 37: 135-139

1513 Bachliński R, Bagiński B (2007) Kłodzko-Złoty-Stok granitoids massif. In
1514 Kozłowski A., Wiszniewska J (eds) In Granitoids in Poland, *Archiwum*
1515 *mineralogiczne*, Monograph 1: 261-273

1516 Baranowski Z, Haydukiewicz A, Kryza R, Lorenc S, Muszynski A (1990) Outline
1517 of the geology of the Góry Kaczawskie (Sudetes, Poland). *Neues*
1518 *Jahrbuch für Geologie und Paläontologie*, Abhandlungen 179: 223-257

1519 Barth SR (2000) Geochemical and boron, oxygen and hydrogen isotopic
1520 constraints on the origin of salinity in groundwaters from the
1521 crystalline basement of the Alpine Foreland. *Applied Geochemistry*
1522 15:937-952. [https://doi.org/10.1016/S0883-2927\(99\)00101-8](https://doi.org/10.1016/S0883-2927(99)00101-8)

1523 Beauvais A, Chardon F (2013) Modes, tempo, and spatial variability of
1524 Cenozoic cratonic denudation: The West African example, *Geochem.*
1525 *Geophys. Geosyst.*, 14, 1590-1608, doi:10.1002/ggge.20093

1526 Beauvais, A., Ruffet G, Henocque O, Colin F (2008) Chemical and physical
1527 erosion rhythms of the West African Cenozoic morphogenesis: The ³⁹Ar-
1528 ⁴⁰Ar dating of supergene K-Mn oxides, *J. Geophys. Res.*, 113, F04007,
1529 doi:10.1029/2008JF000996

1530 Ben Baccar M, Fritz B, Madé B (1993) Diagenetic albitization of K-feldspar
1531 and plagioclase in sandstone reservoirs: thermodynamic and kinetic
1532 modeling. *Journal of Sedimentary Petrology* 63:1100-1109

1533 Besse J, Courtillot V (2003) Apparent true polar wander and the geometry of
1534 the geomagnetic field over the last 200 Myr: Correction. *Journal of*
1535 *Geophysical Research* 108: 2300

1536 Bonhomme M, Yerle J.-J, Thiry M (1980) Datation K-Ar de fractions fines
1537 associées aux minéralisations. Le cas du bassin uranifère permo-
1538 houiller de Brousse-Broquiès (Aveyron, France).- *C.R. Académie*
1539 *Sciences Paris* 291D: 121-123

1540 Boone GM (1969) Origin of clouded red feldspars, petrology contrasts in a
1541 granitic porphyry intrusion. *American Journal of Science* 267: 633-668

1542 Borkowska M, (1966) Petrografia granitu Karkonoszy (Petrography of the
1543 Karkonosze Granite). *Geologia Sudetica* 2: 7-119. (in Polish with
1544 English extended summary)

- 1545 Botor D, Dunkl I, Anczkiewicz A, Mazur S (2017) Post-Variscan thermal
1546 history of the Moravo-Silesian lower Carboniferous Culm Basin (NE
1547 Czech Republic - SW Poland). *Tectonophysics* 712-713: 643-662
- 1548 Boulangé B (1984) Les formations bauxitiques latéritiques de Côte d'Ivoire.
1549 *Travaux et Documents ORSTOM* 175, 341 p
- 1550 Boulvais P, Ruffet G, Cornichet J, Mermet M (2007) Cretaceous albitization
1551 and dequartzification of Hercynian peraluminous granite in the
1552 Salvezines Massif (French Pyrénées). *Lithos* 93: 89-106.
- 1553 Bourquin S, Bercovici A, López-Gómez J, Diez J-B, Broutin J, Ronchi A, Amour
1554 F (2011) The Permian-Triassic transition and the onset of Mesozoic
1555 sedimentation at the northwestern peri-Tethyan domain scale:
1556 palaeogeographic maps and geodynamic implications. *Palaeogeography*
1557 *Palaeoclimatology Palaeoecology* 299: 265-280
- 1558 Bowles JFW, Howie RA, Vaughan DJ, Zussman J (2011) Non-Silicates: Oxides,
1559 Hydroxides and Sulphides. *Rock-Forming Minerals*, 2nd edn, Geological
1560 Society of London 5A: 920 p.
- 1561 Boyce AJ, Fulignati P, Sbrana A (2003) Deep hydrothermal circulation in a
1562 granite intrusion beneath Larderello geothermal area (Italy):
1563 constraints from mineralogy, fluid inclusions and stable isotopes.
1564 *Journal of Volcanology and Geothermal Research* 126: 243-262
- 1565 Brinhall G, Dietrich W (1987) Constitutive mass balance relations between
1566 chemical composition, volume, density, porosity, and strain in
1567 metasomatic hydrochemical systems: Results on weathering and
1568 pedogenesis. *Geochimica et Cosmochimica Acta* 51: 567-587
- 1569 Butler RF (1998) *Paleomagnetism: Magnetic domains to geologic terranes.*
1570 Electronic edition, 237 p.
- 1571 Castet S, Dandurand J-L, Schott J, Gout R (1993) Boehmite solubility and
1572 aqueous aluminium speciation in hydrothermal solutions (90-350°C):
1573 Experimental study and modeling. *Geochimica et Cosmochimica Acta* 57:
1574 4869-4888

- 1575 Cathelineau M (1986) The hydrothermal alkali metasomatism effects on
1576 granitic rocks: quartz dissolution and related subsolidus changes.
1577 Journal of Petrology 27: 945-965
- 1578 Chrzastek A., Wojewoda J (2011) Mezozoik południowo-zachodniej Polski -
1579 synklinorium północnosudeckie (Mesozoic of SW Poland, the North
1580 Sudetic Synclinorium). In: Żelaźniewicz, A., Wojewoda, J., Ciężkowski,
1581 W. (eds), *Mezozoik i kenozoik Dolnego Śląska*. Polskie Towarzystwo
1582 Geologiczne, Wrocław, 1-10 (in Polish, English abstract)
- 1583 Clément J-Y, (1986) Minéralogie, pétrologie et géochimie du Permien de
1584 Lodève (Hérault, France) : diagenèse précoce, altération
1585 feldspathisante et mise en place des minéralisations uranifères.
1586 E.N.S.M.P. Mémoire Sciences de la Terre 2: 137 p.
- 1587 Cocherie A, Albarede F (2001) An improved U-Th-Pb age calculation for
1588 electron microprobe dating of 845 monazite. *Geochimica et Cosmochimica*
1589 *Acta* 65:4509-4522, [https://doi.org/10.1016/S0016-7037\(01\)00753-0](https://doi.org/10.1016/S0016-7037(01)00753-0) 846
- 1590 Cocherie A, Legendre O, Peucat J-J, Kouamelan AN (1998) Geochronology of
1591 polygenetic monazites 847 constrained by in situ electron microprobe
1592 Th-U-total lead determination: Implications for lead 848 behaviour in
1593 monazite. *Geochimica et Cosmochimica Acta* 62: 2475-2497,
1594 [https://doi.org/10.1016/S0016-7037\(98\)00171-9](https://doi.org/10.1016/S0016-7037(98)00171-9)
- 1595 Cogné J-P (2003) PaleoMac: a Macintosh™ application for treating
1596 paleomagnetic data and making plate reconstructions. *Geochemistry*
1597 *Geophysics Geosystems* 4: 1007.
- 1598 Dammer D (1995) Geochronology of chemical weathering processes in the
1599 northern and western Australian regolith. PhD thesis, Australian
1600 National University: 214 p.
- 1601 Danišík M, Migoń P, Kuhlemann J, Evans NJ, Dunkl I, Frisch W (2010)
1602 Thermochronological constraints on the long-term erosional history of
1603 the Karkonosze Mts, central Europe, *Geomorphology* 117: 78-89
- 1604 Danišík M, Štěpančíková P, Evans NJ (2012) Constraining long-term
1605 denudation and faulting history in intraplate regions by multisystem
1606 thermochronology: An example of the Sudetic Marginal Fault (Bohemian

1607 Massif, central Europe). *Tectonics* 31: TC2003,
1608 <https://doi.org/10.1029/2011TC003012>

1609 De Jong G., Williams PJ (1995). Giant metasomatic systems formed during
1610 exhumation of mid-crustal Proterozoic rocks in the vicinity of the
1611 Cloncurry fault, northwest Queensland. *Australian Journal of Earth
1612 science* 42: 281-290

1613 De La Roche H (1957) *Eléments pour l'étude de la saussuritisation. Mémoires
1614 de l'Institut Scientifique de Madagascar* 7D: 217-223

1615 Depciuch T (1972) Wiek bezwzględny (K-Ar) granitoidów Kłodzko Złotostockich
1616 i strefy Niemczy. *Kwartalnik Geologiczny* 16: 103-111

1617 Drake H, Tullborg EL, Annersten H (2008) Red-staining of the wall rock and
1618 its influence on the reducing capacity around water conducting
1619 fractures. *Applied Geochemistry* 23: 1898-1920

1620 Dziejczak K, Teisseyre AK (1990) The Hercynian molasse and younger deposits
1621 in the Intra-Sudetic Depression, SW Poland. *Neues Jahrbuch für
1622 Geologie und Paläontologie, Abhandlungen* 179: 285-305

1623 Edel J-B, Schneider J-L (1995) The Late Carboniferous to Early Triassic
1624 geodynamics evolution of the Variscan Europe in the light of magnetic
1625 overprints in Early Permian rhyolites from the northern Vosges
1626 (France) and the central Black Forest (Germany). *Geophysical Journal
1627 International* 122: 858-876

1628 Edel J-B, Düringer P (1997) The apparent polar wander path of the European
1629 plate in Upper Triassic-Lower Jurassic times and the Liassic
1630 intraplate fracturing of the Pangea: New palaeomagnetic constraints
1631 from NW France and SW Germany. *Geophysical Journal International* 128:
1632 331-344

1633 Edel J-B, Aifa T, Jelenska M, Kadzialko-Hofmökler M, Zelazniewicz A (1997)
1634 Réaimantations des formations paléozoïques des Sudètes polonaises et
1635 courbe de dérive des pôles géomagnétiques d'Europe du Carbonifère
1636 moyen au Jurassique moyen. *Comptes Rendus de l'Académie des Sciences -
1637 Séries IIA - Earth and Planetary Science Letters* 325: 479-486

1638 Edel J B, Schulmann K, Lexa O, Lardeaux J-M. (2018) Late Palaeozoic palaeomagnetic and tectonic
1639 constraints for amalgamation of Pangea supercontinent in the European Variscan belt. Earth-
1640 science reviews, 177: 589-612

1641 Engvik AK, Golla-Schinder U, Berndt J, Austrheim H, Putnis A (2009)
1642 Intragranular replacement of chlorapatite by hydroxy-fluor-apatite
1643 during metasomatism. Lithos 112: 236-246

1644 Engvik AK, Putnis A, Gerald JDF, Austrheim H (2008) Albitisation of
1645 granitic rocks: the mechanism of replacement of oligoclase by albite.
1646 The Canadian Mineralogist 46: 1401-1415

1647 Fàbrega C, Parcerisa D, Rossell JM, Gurenko A, Franke C (2017) Predicting
1648 instrumental mass fractionation (IMF) of stable isotope SIMS analyses
1649 by response surface methodology (RSM). J Anal At Spectrom 32:731-748.
1650 <https://doi.org/10.1039/C6JA00397D>

1651 Fàbrega C, Parcerisa D, Thiry M, Franke C, Gurenko A, Gómez-Gras D, Solé J,
1652 Travé A. (2019) Permian-Triassic red-stained albitized profiles in the
1653 granitic basement of NE Spain: evidence for deep alteration related to
1654 the Triassic palaeosurface. International Journal of Earth Sciences
1655 108: 2325-2347

1656 Farmer VC, Lumsdon DG (1994) An assessment of complex formation between
1657 aluminium and silicic acid in acidic solutions. Geochimica et
1658 Cosmochimica Acta 58: 3211-3334

1659 Feist-Burkhardt S, Götz A, Szulc J, Borkhataria R, Geluk M, Haas J, Hornung
1660 J, Jordan P, Kempf O, Michalík J, Nawrocki J, Reinhardt L, Ricken W,
1661 Röhling H-G, Rüffer T, Török Á, Zühlke R (2008) Triassic. In: McCann
1662 (ed.), The Geology of Central Europe, Volume 2, Mesozoic and Cenozoic.
1663 Geological Society of London, pp 749-821

1664 Ferry JM (1979) Reaction mechanisms, physical conditions, and mass transfer
1665 during hydrothermal alteration of mica and feldspar in granitic rocks
1666 from south-central Maine, USA. Contributions to Mineralogy and
1667 Petrology 68:125-13

- 1668 Fiebig J, Hoefs J (2002) Hydrothermal alteration of biotite and plagioclase
1669 as inferred from intragranular oxygen isotope-and cation-distribution
1670 patterns. *European Journal of Mineralogy*, 14: 49-60
- 1671 Fisher RA (Ed.) (1953) Dispersion on a Sphere. *Philosophical Transactions*
1672 of the Royal Society of London. Series A, vol. 217, 295 p.
- 1673 Fisher NI, Lewis T, Embleton BJJ (1987) *Statistical Analysis of Spherical*
1674 *Data*, Cambridge Univ. Press, Cambridge, 329 p.
- 1675 Głuszyński, A., Aleksandrowski, P. (2022) Late Cretaceous-Early Palaeogene
1676 inversion-related tectonic structures at the northeastern margin of
1677 the Bohemian Massif (southwestern Poland and northern Czechia). *Solid*
1678 *Earth*, 13(8): 1219-1242
- 1679 Goldsmith JR, Laves F (1954) The microcline-sanidine stability relations.
1680 *Geochimica et Cosmochimica Acta* 5: 1-19
- 1681 González-Acebrón L, Arribas J, Mas R (2010) Role of sandstone provenance in
1682 the diagenetic albitization of feldspars. A case study of the Jurassic
1683 Tera Group sandstone (Camereros Basin, NE Spain). *Sedimentary Geology*
1684 229: 53-63
- 1685 González-Acebrón L, Götze J, Barca D, Arribas J, Mas R, Perez-Garrido C
1686 (2012) Diagenetic albitization in the Tera Group, Cameros Basin (NE
1687 Spain) recorded by trace elements and spectral cathodoluminescence.
1688 *Chemical Geology* 312:: 148-162
- 1689 Götze J., Krbetschek M.R., Habermann D., Wolf D. 1999: High-resolution
1690 cathodoluminescence studies of feldspar minerals. [In]:
1691 *Cathodoluminescence in geosciences*. Ed.: Pagel M., Barbin V., Blanc
1692 Ph., Ohnenstetter D., Springer-Verlag, Berlin Heidelberg New York:
1693 271-302
- 1694 Götze J, Krbetschek MR, Habermann D, Wolf D (2000) High-resolution
1695 cathodoluminescence studies of feldspar minerals. In:
1696 *Cathodoluminescence in geosciences*. Ed.: Pagel M., Barbin V., Blanc
1697 Ph., Ohnenstetter D., Springer-Verlag, Berlin Heidelberg New York:
1698 245-270

- 1699 Grimaud J-L, Chardon, Metelka V, Beauvais A, Bamba O (2015) Neogene
1700 cratonic erosion fluxes and landform evolution processes from regional
1701 regolith mapping (Burkina Faso, West Africa). *Geomorphology* 241: 315-
1702 330
- 1703 Guy B (1993) Mathematical revision of Korzhinskii's theory of infiltration
1704 metasomatic zoning. *European Journal of Mineralogy* 5: 317-339
- 1705 Hövelmann J, Putnis A, Geisler T, Schmidt BC, Golla-Schindler U (2010) The
1706 replacement of plagioclase feldspars by albite: observations from
1707 hydrothermal experiments. *Contribution to Mineralogy and Petrology*
1708 159: 43-59
- 1709 Jean A, Beauvais A, Chardon D, Arnaud N, Jayananda M, Mathe PE (2020)
1710 Weathering history and landscape evolution of Western Ghats (India)
1711 from $^{40}\text{Ar}/^{39}\text{Ar}$ dating of supergene K-Mn oxides. *Journal of the*
1712 *Geological Society*, 177(3): 523-536
- 1713 Jenkin GRT, Fallick AE, Leake BE (1992) A stable isotope study of
1714 retrograde alteration in SW Connemara, Ireland. *Contributions to*
1715 *Mineralogy and Petrology* 110: 269-288
- 1716 Jokubauskas P, Bagiński B, MacDonald R, Awdankiewicz M (2014) Field trip to
1717 the Kłodzko - Złoty Stok intrusion. *Mineralogia, Mineralogical Society*
1718 *of Poland, Kraków, Special Papers* 42: 127-134
- 1719 Jokubauskas P, Bagiński B, Macdonald R, Krzemińska E (2018) Multiphase
1720 magmatic activity in the Variscan Kłodzko-Złoty Stok intrusion,
1721 Polish Sudetes: evidence from SHRIMP U-Pb zircon ages. *International*
1722 *Journal of Earth Sciences* 107: 1623-1639
- 1723 Kadzialko-Hofmokr M, Kruczyk J, Mazur S, Siemiatkowski J (2003)
1724 Paleomagnetism of the Upper Paleozoic and devonian rocks from the
1725 Kłodzko metamorphic Complex in the West Sudetes (SWPoland): tectonic
1726 implications for the Variscan Belt of Central Europe. *Tectonophysics*
1727 177: 83-99
- 1728 Kelley, D.S., Robinson, P.T., Malpas, J.G. (1992) Processes of brine
1729 generation and circulation in the oceanic crust: fluid inclusion
1730 evidence from the Troodos ophiolite, Cyprus. *Journal of Geophysical*
1731 *Research: Solid Earth*, 97(B6): 9307-9322

- 1732 Kirschvink JL (1980) The least-square line and plane and the analysis of
1733 paleomagnetic data, *Geophys. J. R. Astron. Soc.* 62: 699-718
- 1734 Kloppmann W, Girard J-P, Négrel P (2002) Exotic stable isotope compositions
1735 of saline waters and brines from the crystalline basement. *Chemical*
1736 *Geology* 184: 49-70. [https://doi.org/10.1016/S0009-2541\(01\)00352-7](https://doi.org/10.1016/S0009-2541(01)00352-7)
- 1737 Korzhinskii DS, Oestreich W (1965) *Abriss der metasomatischen Prozesse.*
1738 Akademie-Verlag Berlin: 195 p.
- 1739 Kowalski A (2020) Triassic palaeogeography of NE Bohemian Massif based on
1740 sedimentological record in the Wleń Graben and the Krzeszów
1741 Brachysyncline (SW Poland). *Annales Societatis Geologorum Poloniae* 90:
1742 125 - 148
- 1743 Kozłowski A, Karwowski Ł, Olszyński W (1975) Tungsten-tin-molybdenum
1744 mineralization in the Karkonosze massif. *Acta Geologica Polonica* 25:
1745 415-430
- 1746 Kroner U, Mansy J-L, Mazur S, Aleksandrowski P, Hann HP, Huckriede H,
1747 Lacquement F, Lamarche J, Ledru P, Pharaoh T, Zedler H, Zeh A, Zulauf
1748 G (2008) Variscan Tectonics. In: McCann, T. (ed.): *The Geology of*
1749 *Central Europe.* The Geological Society, London: 599-66
- 1750 Kryza R, Crowley QG, Larionov A, Pin Ch, Oberc-Dziedzic T, Mochnacka K
1751 (2012) Chemical abrasion applied to SHRIMP zircon geochronology: An
1752 example from the Variscan Karkonosze Granite (Sudetes, SW Poland).
1753 *Gondwana Research* 21: 757-767
- 1754 Kryza R, Schaltegger U, Oberc-Dziedzic T, Pin Ch, Ovtcharova M (2014)
1755 Geochronology of a composite granitoid pluton: a high-precision
1756 ID-TIMS U-Pb zircon study of the Variscan Karkonosze Granite (SW
1757 Poland). *International Journal of Earth Sciences* 103: 683-696
- 1758 Kusiak M.A, Williams IS, Dunkley DJ, Konečný P, Słaby E, Martin HM (2014)
1759 Monazite to the rescue: U-Th-Pb dating of the intrusive history of the
1760 composite Karkonosze pluton, Bohemian Massif. *Chemical Geology* 364:
1761 76-92
- 1762 Lee MR, Parsons I (1997) Dislocation formation and albitization in alkali
1763 feldspars from the Shap granite. *American Mineralogist*, 82: 557-570

- 1764 Leichmann J, Broska I, Zachovalová K (2003) Low-grade metamorphic
1765 alteration of feldspar minerals: a CL study. *Terra Nova*, 15: 104-108
- 1766 Lewandowski, M., Werner, T., Nowoz'yn'ski, K., 1997. PDA—a package of
1767 Fortran programs for paleomagnetic data analysis. *Inst. Geophys. Pol.*
1768 *Ac. Sci.* (Personal communication)
- 1769 Lorenc M (1994) Rola magm zasadowych w ewolucji intruzji granitoidowych
1770 (studium porównawcze wybranych masywów hercyńskich). *Geologia Sudetica*
1771 28: 1-130
- 1772 Ludwig KR (2003) User's Manual for Isoplot 3.00 - A Geochronological
1773 Toolkit for Microsoft Excel. 940 Berkeley Geochronological Centre Spec
1774 Publ Special Pu: 25-32
- 1775 McFadden PL, McElhinny MW (1988) The combined analysis of remagnetization
1776 circles and direct observations in paleomagnetism, *Earth Plan. Sci.*
1777 *Letts.* 87: 161-172
- 1778 Machowiak K, Armstrong R (2007) SHRIMP U-Pb zircon age from the Karkonosze
1779 granite. *Mineralogia Polonica Special Papers* 31: 193-196
- 1780 MacQuarrie KTB, Mayer KU, Jin B, Spiessl SM (2010) The importance of
1781 conceptual models in the reactive transport simulation of oxygen
1782 ingress in sparsely fractured crystalline rock. *Journal of contaminant*
1783 *hydrology* 112: 64-76
- 1784 Majzlan J, Grevel KD, Navrotsky A (2003) Thermodynamics of Fe oxides: Part
1785 II. Enthalpies of formation and relative stability of goethite (α -
1786 FeOOH), lepidocrocite (γ -FeOOH), and maghemite (γ -Fe₂O₃). *American*
1787 *Mineralogist* 88: 855-859
- 1788 Marheine D., Kachlik V., Maluski H., Patočka F., Żelaźniewicz A. 2002. The
1789 ⁴⁰Ar/³⁹Ar ages from the West Sudetes (NE Bohemian Massif): constraints
1790 on the Variscan polyphase tectonothermal development. *Geol. Soc. Spec.*
1791 *Publ.*, 201: 133-155
- 1792 Marshall DJ (1988) *Cathodoluminescence of Geological Materials*. Unwin
1793 Hyman, Boston, 146 p.

- 1794 Matte Ph, Maluski H, Rajlich P, Franke W (1990) Terrane boundaries in the
1795 Bohemian Massif: results of large scale Variscan shearing.
1796 Tectonophysics 177: 151-170
- 1797 Matthews A, Goldsmith JR (1984) The influence of metastability on reaction
1798 kinetics involving zoisite formation from anorthite at elevated
1799 pressures and temperatures. American Mineralogist 69: 848-857
- 1800 Mazur S, Aleksandrowski P, Kryza R, Oberc-Dziedzic T (2006) The Variscan
1801 orogen in Poland. Geological Quarterly 50: 89-118
- 1802 Mazur S, Aleksandrowski P, Turniak K, Awdankiewicz M (2007) Geology,
1803 tectonic evolution and late Paleozoic magmatism of the Sudetes-an
1804 overview. In Kozłowski A., Wiszniewska J (eds). In Granitoids in
1805 Poland. Archiwum mineralogiczne Monograph 1: 59-87
- 1806 Migoń P, Lidmar-Bergström K (2001) Weathering mantles and their
1807 significance for geomorphological evolution of central and northern
1808 Europe since the Mesozoic. Earth-Science Reviews 56: 285-324
- 1809 Migoń P, Danišik M (2012) Erosional history of the Karkonosze Granite
1810 Massif - constraints from adjacent sedimentary basins and
1811 thermochronology. Geological Quarterly 56: 440-454
- 1812 Mikulski SZ, Williams IS, Baginski B (2013) Early Carboniferous (Visean)
1813 emplacement of the collisional Kłodzko-Złoty-Stok granitoids (Sudetes,
1814 SW Poland): constraints from geochemical data and zircon U-Pb ages.
1815 International Journal of Earth Sciences 102: 1007-1027
- 1816 Mikulski S.Z, Bagiński B, Dzierżanowski P (2004) The CHIME age calculations
1817 on monazite and xenotime in aplogranite from the Szklarska Poręba
1818 Huta. Miner. Ass. Pol. Spec. Pap., 24: 287-290
- 1819 Milewicz J., 1968. Karbon górny: niecka północnosudecka. In: Budowa
1820 Geologiczna Polski, t.1, cz. 1. Prekambr i paleozoik. Wyd.
1821 Geologiczne, Warszawa: 439-441
- 1822 Milewicz J., 1970. The Cretaceous of the Jerzmanice Graben (Sudetes). Biul.
1823 Inst. Geol., 239: 37-66

- 1824 Morad S, El-Ghaly MAK, Caja MA, Sirat M, Al-Ramadan K, Mansurbeg H (2010)
1825 Hydrothermal alteration of plagioclase in granitic rocks from
1826 Proterozoic basement of SE Sweden. *Geological Journal* 45: 105-116
- 1827 Oelkers EH, Schott J, Devidal JL (1994) The effect of aluminum, pH, and
1828 chemical affinity on the rates of aluminosilicate dissolution
1829 reactions. *Geochimica et Cosmochimica Acta* 58: 2011-2024
- 1830 Parcerisa D, Thiry M, Schmitt J-M (2009) Albitisation related to the
1831 Triassic inconformity in igneous rocks of the Morvan Massif (France).
1832 *International Journal of Earth Science* 99: 527-544
- 1833 Perez RJ, Boles JR (2005) An empirically derived kinetic model for
1834 albitization of detrital plagioclase. *American Journal of Science* 305:
1835 312-343
- 1836 Peulvast J-P, Sales VC, Bétard F, Gunnell Y (2008) Low post-Cenomanian
1837 denudation depths across the Brazilian Northeast: implications for
1838 long-term landscape evolution at a transform continental margin.
1839 *Global and Planetary Change*, 62(1-2): 39-6
- 1840 Pillans B (2007) Pre-Quaternary landscape inheritance in Australia. *Journal*
1841 *of Quaternary Science: Published for the Quaternary Research*
1842 *Association*, 22(5): 439-447
- 1843 Plümper O., Putnis A (2009) The Complex Hydrothermal History of Granitic
1844 Rocks: Multiple Feldspar Replacement Reactions under Subsolvus
1845 Conditions. *Journal of Petrology* 50(5): 967-987
- 1846 Powolny, T., Dumańska-Słowik, M., Anczkiewicz, A. A., & Sikorska-
1847 Jaworowska, M. (2022). Origin and timing of spilitic alterations in
1848 volcanic rocks from Głuszycza Górna in the Intra-Sudetic Basin, Poland.
1849 *Scientific reports*, 12(1), 1-26. [https://doi.org/10.1038/s41598-022-](https://doi.org/10.1038/s41598-022-15644-2)
1850 [15644-2](https://doi.org/10.1038/s41598-022-15644-2)
- 1851 Putnis A, Hinrichs R, Putnis CV, Golla Schindler U, Collins LG (2007)
1852 Hematite in porous red-clouded feldspars; evidence of large-scale
1853 crustal fluid-rock interaction. *Lithos* 95: 10-18
- 1854 Putnis CV, Ruiz-Agüido E (2013) The mineral-water interface: Where minerals
1855 react with the environment. *Elements* 9: 177-182

1856 Ramseyer K, Boles JR, Lichtner PC (1992) Mechanism of plagioclase
1857 albitization. *Journal of Sedimentary Petrology* 62: 349-356

1858 Reicherter K, Froitzheim N, Jarosiński M, Badura J, Franzke H-J, Hansen M,
1859 Hübscher H, Müller R, Poprawa P, Reinecker J, Stackebrandt W, Voigt T,
1860 Von Eynatten H, Zuchiewicz W (2008). Alpine tectonics north of the
1861 Alps. In: McCann (ed), *The Geology of Central Europe, Volume 2:*
1862 *Mesozoic and Cenozoic.* Geological Society, London, pp 1233-1285

1863 Richter L., Diamond LW (2022) Characterization of hydrothermal fluids that
1864 alter the upper oceanic crust to spilite and epidosite: Fluid
1865 inclusion evidence from the Semail (Oman) and Troodos (Cyprus)
1866 ophiolites. *Geochimica et cosmochimica acta*, 319: 220-253

1867 Ricordel C, Parcerisa D, Thiry M, Moreau M-G, Gómez-Gras D (2007) Triassic
1868 magnetic overprints related to albitization in granites from the
1869 Morvan massif (France). *Palaeogeography Palaeoclimatology*
1870 *Palaeoecology* 251: 268-282.
1871 <https://doi.org/10.1016/j.palaeo.2007.04.001>

1872 Ruffet G, Innocent C, Michard A, Féraud G, Beauvais A, Nahon D, Hamelin B
1873 (1996) A geochronological $^{40}\text{Ar}/^{39}\text{Ar}$ and $^{87}\text{Rb}/^{87}\text{Sr}$ study of K-Mn oxides
1874 from the weathering sequence of Azul, Brazil. *Geochimica et*
1875 *Cosmochimica Acta* 60: 2219-2232

1876 Saigal GC, Morad S, Bjørlykke K, Egeberg PK, Aagaard P (1988) Diagenetic
1877 albitization of detrital K-feldspars in Jurassic, Lower Cretaceous,
1878 and Tertiary clastic reservoir rocks from offshore Norway, I Textures
1879 and origin. *Journal of Sedimentary Petrology* 58:1003-1013

1880 Sandström B, Annersten H, Tullborg EL (2010) Fracture-related hydrothermal
1881 alteration of metagranitic rock and associated changes in mineralogy,
1882 geochemistry and degree of oxidation: a case study at Forsmark,
1883 central Sweden. *International Journal of Earth Science* 99: 1-25

1884 Schmitt J-M (1986) Albitisation triasique, hydrothermalisme jurassique et
1885 altération supergène récente: métallogénie des gisements uranifères du
1886 Rouergue. *Doct. ès Sciences Thesis, Strasbourg, Louis Pasteur*
1887 *University: p 240.*

- 1888 Schmitt J-M (1992) Triassic albitization in southern France: an unusual
1889 mineralogical record from a major continental paleosurface. In:
1890 Schmitt JM, Gall Q (eds) Mineralogical and geochemical records of
1891 paleoweathering. ENSMP, Mémoires des Sciences de la Terre, Paris, pp
1892 115-131
- 1893 Schmitt J-M (1994) Geochemical modeling and origin of the Triassic
1894 albitized regolith in Southern France. 14th International
1895 Sedimentological Congress, Aug. 20-28, Recife, Brazil, Abstracts S8,
1896 p. 19-21
- 1897 Schmitt J-M Baubron JC, Bonhomme MG (1984) Pétrographie et datations K-Ar
1898 des transformations minérales affectant le gîte uranifère de
1899 Bertholène (Aveyron-France). Mineralium Deposita 19:123-131
- 1900 Schmitt J-M, Clement JY (1989) Triassic regolithization: a major stage of
1901 pre-enrichment in the formation of unconformity related deposits in
1902 Southern France. Metallogenesis of uranium deposits. IAEA Technical
1903 Committee Meeting, Vienna 542(8): 93-113.
- 1904 Simon K, Hoefs J (1987) Effects of meteoric water interaction on Hercynian
1905 granites from the Südschwarzwald, southwest Germany. Chemical Geology
1906 61: 253-261
- 1907 Słaby E, Götze J (2004) Feldspar crystallization under magma-mixing
1908 conditions shown by cathodoluminescence and geochemical modelling—a
1909 case study from the Karkonosze pluton (SW Poland). *Mineralogical
1910 Magazine*, 68(4), 561-577
- 1911 Słaby E, Martin H (2008) Mafic and felsic magma interaction in granites:
1912 the Hercynian Karkonosze pluton (Sudetes, Bohemian massif). *Journal of
1913 Petrology* 49: 353-391
- 1914 Słaby E, Galbarczyk-Gąsiorowska E, Baszkiewicz A (2002) Mantled alkali-
1915 feldspar megacrysts from the marginal part of the Karkonosze granitoid
1916 massif (SW Poland). *Acta Geologica Polonica* 52: 501-519
- 1917 Słaby E, Galbarczyk-Gąsiorowska E, Seltmann R, Müller A (2007) Alkali
1918 feldspar megacryst growth: Geochemical modelling. *Mineralogy and
1919 Petrology* 89: 1-29

- 1920 Sobczyk A, Danišík M, Wojewoda J, Śliwiński W, Raczyński P, August C (2013)
1921 Zircon (U-Th)/He thermochronology of the Permian Weissliegende
1922 sandstone in the Fore-Sudetic Homocline (SW Poland). *Mineralogia -*
1923 *Special Papers*: 41: 81.
- 1924 Sobczyk A, Danišík M, Aleksandrowski P, Anczkiewicz A (2015) Post-Variscan
1925 cooling history of the central Western Sudetes (NE Bohemian Massif,
1926 Poland) constrained by apatite fission-track and zircon (U-Th)/He
1927 thermochronology. *Tectonophysics* 649: 47-57
- 1928 Sobczyk A, Sobel ER, Georgieva V (2020) Meso-Cenozoic cooling and
1929 exhumation history of the Orlica-Śnieżnik Dome (Sudetes, NE Bohemian
1930 Massif, Central Europe): Insights from apatite fission-track
1931 thermochronometry. *Terra Nova* 32 (2): 122-133
- 1932 Steinthorsson S, Helgason O (1992) Maghemite in Icelandic basalts.
1933 *Mineralogical Magazine* 56: 185-199
- 1934 Stillings LL, Brantley SL (1995) Feldspar dissolution at 25°C and pH 3:
1935 Reaction stoichiometry and the effect of cations. *Geochimica et*
1936 *Cosmochimica Acta* 59: 1483-1496
- 1937 Stober I, Bucher K (1999a) Origin of salinity of deep groundwater in
1938 crystalline rocks. *Terra Nov* 11:181-185.
1939 <https://doi.org/10.1046/j.1365-3121.1999.00241.x>
- 1940 Stober I, Bucher K (1999b) Deep groundwater in the crystalline basement of
1941 the Black Forest region. *Applied Geochemistry* 14: 237-254.
1942 [https://doi.org/10.1016/S0883-2927\(98\)00045-6](https://doi.org/10.1016/S0883-2927(98)00045-6)
- 1943 Sun SS, Eadington PJ (1987) Oxygen isotope evidence for the mixing of
1944 magmatic and meteoric waters during tin mineralization in the Mole
1945 Granite, New South Wales, Australia. *Economic Geology* 82: 43-52
- 1946 Suzuki K, Adachi M (1994) Middle Precambrian detrital monazite and zircon
1947 from the hida gneiss on Oki-Dogo Island, Japan: their origin and
1948 implications for the correlation of basement gneiss of Southwest Japan
1949 and Korea. *Tectonophysics* 235: 277-292. [https://doi.org/10.1016/0040-](https://doi.org/10.1016/0040-1951(94)90198-8)
1950 [1951\(94\)90198-8](https://doi.org/10.1016/0040-1951(94)90198-8)

- 1951 Suzuki K, Adachi M, Tanaka T (1991) Middle precambrian provenance of
1952 Jurassic sandstone in the Mino Terrane, central Japan: Th-U-total Pb
1953 evidence from an electron microprobe monazite study. *Sedimentary*
1954 *Geology* 75: 141-147. [https://doi.org/10.1016/0037-0738\(91\)90055-I](https://doi.org/10.1016/0037-0738(91)90055-I)
- 1955 Suzuki K, Adachi M, Kajizuka I (1994) Electron microprobe observations of
1956 Pb diffusion in metamorphosed detrital monazites. *Earth and Planetary*
1957 *Science Letters* 128: 391-405. [https://doi.org/10.1016/0012-1079-](https://doi.org/10.1016/0012-1079-821X(94)90158-9)
1958 [821X\(94\)90158-9](https://doi.org/10.1016/0012-1079-821X(94)90158-9)
- 1959 Taylor HP (1977) Water/rock interactions and the origin of H₂O in granitic
1960 batholiths. *Journal of Geological Society of London* 133: 509-558
- 1961 Taylor HP (1977) Water/rock interactions and the origin of H₂O in granitic
1962 batholiths: Thirtieth William Smith lecture. *J Geol Soc London* 133:509-
1963 558. doi: 10.1144/gsjgs.133.6.0509
- 1964 Thiry M, Liron MN, Dubreucq P, Polton J-C (2017) *Curiosités géologiques du*
1965 *massif de Fontainebleau, Guide géologique, BRGM Editions., 115 p.*
- 1966 Thiry M, Milnes AR, Rayot V, Simon-Coinçon R (2006) Interpretation of
1967 palaeoweathering features and successive silicifications in
1968 the Tertiary regolith of Inland Australia. *Journal of the Geological*
1969 *Society* 163: 723-736
- 1970 Vasconcelos PM, Becker TA, Renne PR, Brimhall GH (1994) Direct dating of
1971 weathering phenomena by ⁴⁰Ar/³⁹Ar and K-Ar analysis of supergene K-Mn
1972 oxides. *Geochimica et Cosmochimica Acta* 58: 1635-1665
- 1973 Vercruysse C (2011) *Mise en évidence et datation de la paléosurface*
1974 *crétacée dans le massif des Vosges. Rapport No 0110620CVER, Centre de*
1975 *Geosciences, Ecole des Mines de Paris, Fontainebleau, France, 75 p.*
- 1976 Waychunas GA (1991) *Oxide Minerals: Petrologic and magnetic significance.*
1977 *Reviews in Mineralogy* 25: 38-46.
- 1978 Wierzchołowski B (1976) *Granitoidy kłodzko-złotostockie i ich kontaktowe*
1979 *oddziaływanie na skały osłony (studium petrograficzne). Geologia*
1980 *Sudetica* 11: 7-147

- 1981 Wilamowski A, (2002) Chloritization and polytypism of biotite in the
1982 Łomnica granite, Karkonosze Massif, Sudetes, Poland: Stable isotope
1983 evidence. *Chemical Geology* 182(2-4): 529-547
- 1984 Williams ML, Jercinovic MJ, Goncalves P, Mahan K (2006) Format and
1985 philosophy for collecting, 1096 compiling, and reporting microprobe
1986 monazite ages. *Chemical Geology* 225: 1-15. <https://doi.org/10.1016/j.mce.2005.09.009>
1987
- 1988 Wojdyr M (2010) Fityk: A general-purpose peak fitting program. *Journal of*
1989 *Applied Crystallography* 43: 1126-1128
- 1990 Yao KFE (2013) Albitization and oxidation of the granitoid rocks related to
1991 the Triassic paleosurface in the Sudetes (SW Poland). Thesis École
1992 Nationale Supérieure des Mines de Paris in Paris (France) and
1993 Państwowy Instytut Geologiczny - Państwowy Instytut Badawczy in Warsaw
1994 (Poland). 164 p. <http://pastel.archives-ouvertes.fr/pastel-00971314>
- 1995 Yerle J.-J, Thiry M (1979) Albitisations et minéralisations uranifères dans
1996 le socle et les sédiments permo-houillers du bassin de Brousse-
1997 Broquiès (Aveyron, France). *Bulletin BRGM* 4: 275-290
- 1998 Žák J, Klominsky J (2007) Magmatic structures in the Krkonoše-Jizera
1999 Plutonic Complex, Bohemian Massif: evidence for localized multiphase
2000 flow and small-scale thermal-mechanical instabilities in a granitic
2001 magma chamber. *Journal of Volcanology and Geothermal Research* 164:
2002 254-267
- 2003 Žák J, Verner K, Sláma J, Kachlík V, Chlupáčová M (2013) Multistage magma
2004 emplacement and progressive strain accumulation in the shallow-level
2005 Krkonoše-Jizera plutonic complex, Bohemian Massif. *Tectonics* 32: 1493-
2006 1512
- 2007 Zheng Y-F (1993) Calculation of oxygen isotope fractionation in
2008 anhydrous silicate minerals. *Geochim Cosmochim Acta* 57:1079-
2009 1091. doi: 10.1016/0016-7037(93)90042-U
- 2010 Zijderfeld, JDA. (1967) A.C. demagnetisation of rocks: analysis of
2011 results. In: Collinson, DW, Creer, KM, Runcorn, SK (Eds.),
2012 *Methods in Paleomagnetism*. Elsevier, Amsterdam: 254-286