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1 **Jolts in the Jade Factory:**
2 **a Route for Subduction Fluids and**
3 **their Implications for Mantle Wedge Seismicity**

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20

21

22 **Keywords**

23 Subduction, Fluids, Jadeitite, Earthquakes, Mantle wedge, Foliated cataclasites

24 **Abstract**

25 An increasing number of seismological studies report transient seismicity clusters in the mantle wedge
26 several kilometers above the subduction interface. Their physical significance with respect to subduction
27 zone seismo-tectonics remains poorly understood. Jadeitites are known to form and/or be associated
28 with mantle wedge serpentinites in the c. 30-70 km depth range, and thus may yield information on
29 deformation mechanisms in this region of deep subduction environments. We herein document and
30 compare brittle-viscous features recorded in jadeitites from Polar Urals (Russia), Kashin state (Myanmar)
31 and Motagua fault region (Guatemala) – some of the most important jadeitite occurrences worldwide. In
32 the Polar Urals we identified ultramafic-hosted pristine jadeitite-bearing veins c. 1 km above a Devonian
33 paleo-subduction interface, interpreted as metasomatized former felsic dyke networks crosscutting the
34 mantle wedge peridotites. Here, both jadeitites and associated amphibole-rich dark granofels display
35 widespread brittle-ductile deformation fabrics such as shear bands, foliated cataclasites and breccias,
36 cemented through dissolution-precipitation processes by omphacite and sodic amphiboles, a mineral
37 assemblage typical of high-pressure–low-temperature subduction zone conditions. Electron probe and
38 laser ablation ICP-MS mapping indicate that these brittle-viscous networks display a substantial
39 metasomatic imprint highlighted in the dark granofels by variations in major and trace elements.
40 Switches between viscous and brittle deformation patterns are attested by crystallographic-preferred
41 orientations of jadeite in some of the shear zones that crosscut the host jadeitites. Strikingly similar
42 mineral assemblages and deformation patterns were observed in the Kashin and Motagua samples.
43 Observed deformation features in these localities can be classified into three categories (tectonic
44 breccias, foliated cataclasites and hydraulic breccias), which may occasionally form in sequence and
45 exhibit mutually overprinting textures. Some of the foliated cataclasites contain fine-grained and foliated
46 “shard-like” features forming a radial omphacite-jadeite spherulitic texture, interpreted as former
47 pseudotachylyte that evokes a paleo-seismic origin. We interpret these healed fault networks as

48 recording external fluid influx within fracture zones that repeatedly ruptured along former “dyke”
49 networks. These high permeability drains likely (i) contribute to the transfer of highly pressurized plate-
50 interface metamorphic fluids into the mantle wedge; and (ii) trigger seismic instabilities recorded in the
51 basal part of active mantle wedge sections. These findings provide new insights into the current
52 understanding of the rheology (e.g., serpentinization ratio) and stress state in the mantle wedge, with
53 implications for subduction interface seismogenesis.

54 **1. Introduction**

55 Understanding processes rooted in the mantle wedge region of subduction zones is of critical
56 importance because its structure strongly controls the rheology of the subduction interface at the
57 downdip end of the seismogenic zone (Hyndman & Peacock, 2003; Dessa et al., 2009; Agard et al., 2018).
58 Fluids released by metamorphic reactions in the downgoing plate are thought to control the degree of
59 hydration (serpentinization) of the fore-arc mantle wedge (e.g., Hacker et al., 2003; Deschamps et al.,
60 2010; Bostock, 2013). The bulk serpentinization ratio of the mantle wedge, estimated based on the
61 perturbation of seismic wave velocities, is generally considered low (i.e., <20 vol.%; Abers et al., 2017)
62 except for specific subduction environments where it may reach up to 60 vol.% (e.g., Central Japan:
63 Hyndman & Peacock, 2003; Mariana margin: Hussong, 1981). The serpentinized mantle wedge has long
64 been considered as mostly aseismic due to the presence of weak minerals such as antigorite or talc that
65 are known to substantially reduce rock strength (e.g., Hilairet et al., 2007). Yet, insights gained from
66 detailed studies on the Sumatra 2004 ($M_w = 9.1$) and Maule 2010 ($M_w = 8.8$) mega-earthquakes have
67 shown that some of these exceptional rupture events may nucleate or propagate at the base of the “cold
68 nose” region (e.g., Dessa et al., 2009; Wang et al., 2020). This paradox raises important questions
69 regarding the rheology of the partly serpentinized mantle above the subduction interface. The report of
70 slow earthquakes near the seismogenic downdip end of several subduction megathrusts also changed
71 our vision regarding stress distribution along the deep interface (e.g., Fu & Freymuller, 2013; Frank et al.,

72 2015; Audet & Kim, 2016), with potential implications for mega-earthquake prediction (Obara & Kato,
73 2016; Bouchon et al., 2018). Whereas it is now accepted that fluids dramatically impact the mechanical
74 stability of the deep serpentinized interface, very little *in situ* information is known about fluid-rock
75 interaction processes or the feedback between upward transported fluids and seismicity (e.g., Angiboust
76 et al., 2014; Locatelli et al., 2018).

77 Over the last decade, an increasing number of high-resolution seismological studies have
78 identified the presence of seismicity nests in the partly serpentinized mantle wedge (e.g., Halpaap et al.,
79 2019 and references therein). These events are generally interpreted as mechanically related to the
80 influx of fluids or melts between 30 and 70 km depth (e.g., Davey & Ristau, 2011). A majority of the
81 clusters identified in literature concentrates in the first 15 km above the plate interface (e.g., Greece:
82 Halpaap et al., 2019; Japan: Nakajima & Uchida, 2018; New Zealand: Davey & Ristau, 2011; Central Chile:
83 Wang et al., 2020; Colombia: Chang et al., 2019) along steeply-dipping, planar features commonly seen
84 as “vent-like” structures, apparently channelizing plate-interface fluids towards the inner wedge.
85 Normal, strike-slip and thrust focal mechanisms are reported for a large majority of these supra-slab
86 earthquakes, with magnitudes generally spanning a range between 2 and 5 (e.g., Halpaap et al., 2019).
87 The physics of the rupture and the nature of the material where these earthquakes are nested remain
88 unknown. Moreover, the source and the composition of the fluids passing through these networks is also
89 a matter of discussion since the precise location of hydrous mineral breakdown reactions strongly
90 depends on the subduction thermal structure (e.g., Hermann et al., 2006; Syracuse et al., 2010).

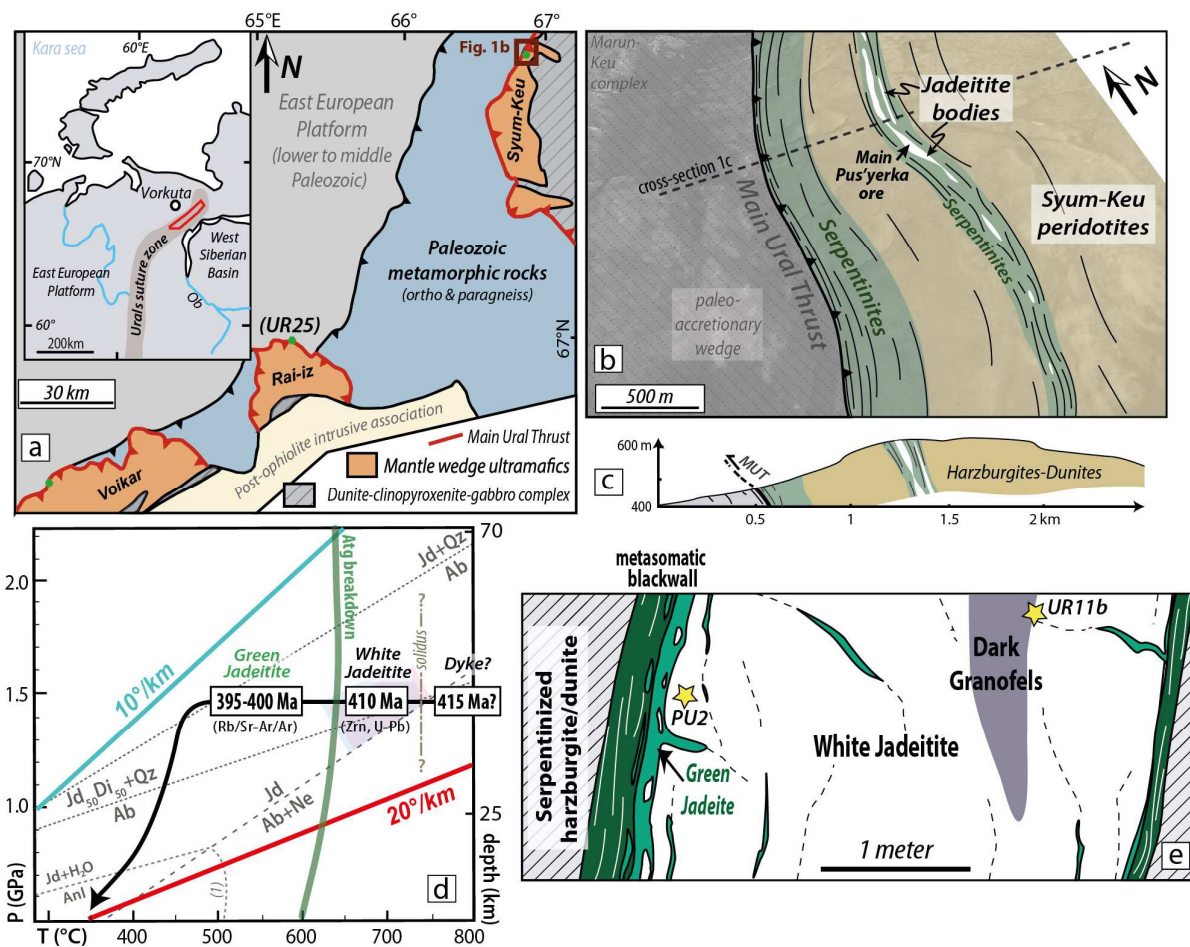
91 Natural mantle wedge samples, despite their scarcity worldwide, represent a unique opportunity
92 to shed light on these deep-seated processes (e.g., Kepezhinskas et al., 1995; Horn et al., 2020).
93 Jadeitites are known to represent fossilized fluid pathways from the base of the hydrated mantle wedge
94 (Harlow & Sorensen, 2005; Harlow et al., 2015). From this perspective, they may provide insights on the
95 physical nature of supra-slab seismic events. However, primary structures from jadeitite-bearing

96 localities have been almost systematically overprinted during long-term subduction, extensive
97 serpentinization and exhumation (e.g., Central America: Flores et al., 2015; Kawamoto et al., 2018;
98 Myanmar: Shi et al., 2009a; Japan: Morishita et al., 2007). In the Polar Urals (Russia), the Pus'yerka
99 locality exhibits a relatively undisturbed contact between a jadeitite "dyke" network and its ultramafic
100 host (e.g., Meng et al., 2011; Angiboust et al., 2021), thus providing an opportunity to identify
101 deformation processes rooted in these jadeitite bodies. Through a combined petrological,
102 microstructural and geochemical investigation of Polar Urals samples, we provide new evidence for
103 brittle-ductile switches in jadeitites (and associated amphibole-phlogopite granofels) microstructures.
104 These structures are compared with those from samples of loose jadeitite boulders from Myanmar and
105 Guatemala, settings where pristine structures are only exceptionally exposed (e.g., Sorensen et al.,
106 2010). We then evaluate their potential meaning in terms of fluid pathways and the genesis of seismic
107 instabilities in the basal region of active mantle wedges.

108 **2. Geological setting**

109 *2.1 Pus'yerka jadeitite deposit (Polar Urals, Russia)*

110 The Polar Urals belt formed during closure of the Uralian ocean by subduction and by the
111 eastward burial of the European continental margin under an oceanic volcanic arc (e.g., Savelieva et al.,
112 2002). In the Polar Urals, the Main Ural Thrust (MUT; **Fig.1a**) corresponds to a major crustal-scale shear
113 zone with peridotites that were thrust over eclogitized continental crust (Marun-Keu complex; e.g.,
114 Udovkina, 1971; Dobretsov & Sobolev, 1984; Glodny et al., 2003). Rare ophiolitic, blueschist-facies
115 mélangé exposures are restricted to the base of the MUT (e.g., Kazak et al., 1976). Locally, the MUT has
116 been subject to moderate reactivation as a detachment fault during exhumation (e.g., Sychev & Kulikova,
117 2012). In the Polar Urals,



118

119 **Figure 1.** a. Simplified geological map of the Polar Urals locating the three main mantle wedge sections
 120 exposed in this region and the study area at the base of the Syum-Keu massif. The inset localizes the Polar
 121 Urals in northern Russia. Green dots correspond to jadeitite-bearing localities (modified after Angiboust et
 122 al., 2021). b. Geological map showing the structures above the Main Ural Thrust, interpreted as an ancient
 123 subduction interface. The studied jadeitite “dyke” crops out as a boudinaged sliver wrapped by a network
 124 of sheared serpentinites, approximately 1 km above the Main Ural Thrust. c. Cross-section highlighting the
 125 geometry of the structures identified in the field (modified after Angiboust et al., 2021). d. Pressure-
 126 Temperature-time sketch summarizing the long-term evolution of the Pus’yerka jadeitite dyke structure
 127 (after Angiboust et al., 2021). $Jd_{50}Di_{50}$ reaction line is after Maruyama & Liou (1988). Reaction (1):
 128 Analcime = Albite + Nepheline. e. Schematic view of the white jadeitite “dyke” as observed in situ in the
 129 Polar Urals with dismembered patches of dark granofels and dark amphibole-phlogopite bearing
 130 blackwalls forming at the contact with the host serpentinitized ultramafic rocks. Mineral abbreviations after
 131 Whitney & Evans (2010).

132

133 three peridotite massifs (the Rai-Iz, Syum-Keu and Voikar massifs) were locally transformed into
134 antigorite-schists along the MUT hangingwall (**Fig.1b**). These large mantle exposures, mostly harzburgitic
135 and lherzolitic in composition (Savelieva & Suslov, 2014; Shmelev et al., 2011), exhibit large chromite
136 deposits and are crosscut by numerous subvertical dunitic channels interpreted as the melt extraction
137 pathways (Batanova et al., 2011). The MUT hanging wall displays (i) a serpentinization gradient towards
138 the underlying unit; (ii) the presence of jadeite ($\text{NaAlSi}_2\text{O}_6$) veins in the serpentinites from the hanging
139 wall; and (iii) a high pressure–low temperature (HP-LT) metamorphic imprint in the footwall units (e.g.,
140 Dobretsov & Ponomareva, 1968; Glodny et al., 2003; Batanova et al., 2011; Shmelev, 2011; Meng et al.,
141 2016). The Rai-Iz, Syum-Keu and Voikar peridotite massifs in the Polar Urals can thus be viewed as good
142 analogues of a supra-subduction setting, enabling the understanding of deep-seated processes below an
143 inferred Paleozoic island arc (e.g., Batanova et al., 2011; Savelieva et al., 2002, 2016; Angiboust et al.,
144 2021), in a region of the deep subduction interface that is not commonly exhumed (Guillot et al., 2009;
145 Agard et al., 2018).

146 The studied Pus'yerka exposure is located along a jadeitite-bearing serpentinized shear zone that
147 is several hundred meters thick and 5 km long, striking N-S and dipping E approximately 1 km above the
148 Main Ural Thrust (MUT; **Fig.1b**). In a recent study, Angiboust et al. (2021) have interpreted the MUT
149 hanging wall as a rare natural case study highlighting the structure along the base of a subduction mantle
150 wedge. This major jadeitite deposit, discovered and mined in the 1980s (Kuznetsov et al., 1986; Fishman,
151 2006), represents a unique locality to investigate the structural contacts between the jadeite bodies and
152 its host (e.g., Meng et al., 2011; Angiboust et al., 2021).

153 The basal serpentinites as well as the jadeitite-bearing network exhibit a regional foliation parallel
154 with the MUT (**Fig.1c**). Field, geochemical and density measurements indicate serpentinization ratios in
155 the range of 35-65 vol.% (in agreement with the 45-65 vol.% estimates from Makeyev, 1992), with up to

156 90-100% approaching the MUT and within the jadeite-bearing networks (**Fig.1b**; Angiboust et al., 2021
157 and references therein). The serpentinite schists that host the jadeitite boudins are composed of
158 antigorite with minor amounts of brucite, magnetite and phlogopite (Makeyev, 1992) as well as chlorite,
159 tremolite and magnesite. Within the serpentinite shear zone, tens of whitish jadeitite lenses are
160 observed, elongated *en echelon* and stretched parallel to the main foliation dip direction (**Fig.1b**). The
161 thickness of these lenses ranges between several tens of centimeters to several meters in the few places
162 where the lenses were observed *in situ* (e.g., Kuznetsov et al., 1986; Meng et al., 2011).

163 Ion probe U-Pb dating of zircon crystals from the main jadeitite body yielded U-Pb ages of 404 ± 7
164 Ma (Meng et al., 2011) and 409 ± 3.3 Ma (Konovalov & Sergeev, 2015). These Devonian ages were
165 interpreted by Meng et al. (2011) as dating intra-oceanic subduction initiation of the Uralian ocean
166 realm. Angiboust et al. (2021) proposed that these jadeitite bodies derive from the metasomatic
167 replacement of a former trondhjemitic dyke that crystallized from slab-derived melts within an
168 ultramafic mantle wedge setting, in a subduction initiation context (i.e., at a temperature regime much
169 higher than expected in a long-lived subduction context; e.g., Soret et al., 2016 and references therein).
170 Ar-Ar plateau ages and multi-mineral Rb-Sr dating yield ages for phlogopite and amphibole-bearing
171 domains ranging from 410–395 Ma, interpreted by Angiboust et al. (2021), in line with the pioneering
172 study of Dobretsov & Ponomareva (1968), as marking the re-equilibration of dyke material during secular
173 cooling from supra-solidus to HP-LT conditions (i.e., from $T > 700^\circ\text{C}$ to $T < 500^\circ\text{C}$ for a pressure on the
174 order of 1.5 GPa). The studied samples were collected in the main jadeitite “dyke” (**Fig.1b,e**).

175 2.2 Kashin state Jade Mines Belt (northern Myanmar)

176 One of the world’s largest jadeitite-bearing suture zones is exposed in the Kashin state of
177 Myanmar in the famous Jade Mines Belt (e.g., Shi et al., 2012; Nyunt et al., 2017), where loose jadeitite
178 fragments are found in conglomerates, river beds or exceptionally embedded within strongly-weathered

179 antigorite schists (e.g., Harlow et al., 2015 and references therein; Ridd et al., 2019). The original
180 jadeitite-bearing structures (likely ancient felsic dykes; e.g., Bleeck, 1908) were formed within a
181 serpentinized mantle wedge from a subduction zone of debated Late Jurassic to Late Cretaceous age
182 (Goffé et al., 2002; Shi et al., 2009a, 2012; Yui et al., 2012; Harlow et al., 2015). These metasomatized
183 dykes were extensively affected by exhumation and subsequent strike-slip deformation related to the
184 Sagaing transform fault system (Harlow et al., 2015; Searle et al., 2017; Ridd et al., 2019). Protracted
185 deformation led to the formation of a serpentinite mélange in which jadeitite “dykes” and blocks were
186 disrupted and disseminated in the antigorite schist matrix, together with other lenses of seafloor origin
187 such as graphite schists, glaucophane schists, garnet amphibolites and garnet micaschists (Nyunt et al.,
188 2017).

189 Although the original thickness of the jadeitite “dykes” is challenging to evaluate due to poor
190 exposure conditions and late deformation, some studies mention typical thicknesses on the order of
191 several meters (Shi et al., 2012, Harlow et al., 2015 and references therein), in line with the structures
192 observed *in situ* in the Polar Urals (**Fig.1e**). Texturally secondary chlorite schists (“blackwalls”) as well as
193 albitite bands are reported at the contact between the dyke structure and the host (Bleeck, 1908;
194 Chhibber, 1934). Na-amphibole-rich bands (mostly eckermannite and glaucophane; see Oberti et al.,
195 2015 for further details on mineralogy) are also found either embedded within the jadeitite “dyke” or
196 lining the contact with the ultramafic host (Bleeck, 1908; Nyunt, 2009). Most pressure-temperature
197 estimates for jadeitite formation in the Jade Mine Belt region span a wide range from 1.0–1.5 GPa and
198 300–500°C (Mével & Kiénast, 1986; Goffé et al., 2002; Shi et al., 2003). The herein studied samples,
199 provided by a local miner, were found as boulders in a conglomerate near the Lonkin township (near
200 Hpakan).

201 2.3 Motagua fault zone (Guatemala)

202 The Guatemala suture zone is an E-W-trending major plate boundary zone that separates the
203 Caribbean and North American plates. This strike-slip suture, which puts in contact the Maya block to the
204 north with the Chortis block to the south (e.g., Ortega-Gutierrez et al., 2007 and references therein),
205 contains many mafic and ultramafic blocks as well as eclogite-, blueschist- and garnet amphibolite-facies
206 crustal and sedimentary fragments interpreted as remnants from a Cretaceous meta-ophiolite (e.g.,
207 Brueckner et al., 2009). The Motagua fault zone separates two distinct terranes likely exhumed during
208 two distinct collisional events: the North Motagua and the South Motagua mélanges (e.g., Beccaluva et
209 al., 1995; Gendron et al., 2002; Harlow et al., 2004; Harlow, 2011). Different metamorphic ages on
210 jadeitites and eclogites were reported for these two terranes, ranging from c. 100–60 Ma for the North
211 Motagua and c. 160–110 Ma for the South Motagua mélanges (see Flores et al., 2013 and references
212 therein). The highest-pressure rocks from these two mélanges are also slightly different, with 500–650 °C
213 and 1.5–2.3 GPa for the North Motagua mélange (Harlow et al., 2008; Tsujimori et al., 2004) and 470–
214 520 °C and 2.0–2.7 GPa (Tsujimori et al., 2006; Endo et al., 2012) for the South Motagua mélange.

215 Jadeitites occur as meter-sized blocks within the serpentinite mélanges but are most commonly
216 found as pebbles within streams or as loose blocks in slope debris. As for the Jade Mines Belt in
217 Myanmar, pristine tectonic relationships are extremely difficult to document. The mineralogy of
218 jadeitites is quite varied in terms of minerals and mineral abundances, and with contrasting assemblages
219 in the north and south Motagua mélanges (Harlow et al., 2011; Flores et al., 2013). The herein studied
220 samples were collected as loose boulders in streams from the southern Motagua mélange (Rio El Tambor
221 area) and contain essentially jadeite plus minor omphacite and lawsonite, in agreement with the
222 mineralogy of jadeitites from this mélange (Harlow et al., 2011).

223 **3. Analytical methods**

224 *3.1 Electron probe microanalysis*

225 Mineral compositions were quantified via electron probe microanalysis (EPMA) using a Cameca
226 SXFive operated in the CAMPARIS analytical facility at Paris University. Standard analytical conditions (15
227 keV, 10 nA, beam diameter 5 μm) and a set of synthetic and natural crystals for calibration standards
228 were used: Fe_2O_3 (Fe), MnTiO_3 (Mn, Ti), diopside (Mg, Si), CaF_2 (F), orthoclase (Al, K), anorthite (Ca) and
229 albite (Na). X-ray maps were acquired on the same instrument using analytical conditions of 15 keV, 250
230 nA, a dwell time of 60 milliseconds and a step size of 2 μm . Some of the X-ray images were processed
231 and quantified with DWImager software (Torres-Roldán and Garcia-Casco, unpublished; see Garcia-
232 Casco, 2007). A scanning electron microscope (SEM) Zeiss EVO MA10 at the Institut de Physique du
233 Globe de Paris using internal calibration standards was used for microscopic investigations, energy
234 dispersive X-ray spectral (EDS) mapping and surface composition characterizations. Mineral
235 abbreviations are from Whitney & Evans (2010). Clinopyroxene and clinoamphibole compositions,
236 including estimation of Fe^{3+} , are calculated according to the schemes of Morimoto (1989) and
237 Hawthorne & Oberti (2007), respectively. Classification of these minerals also follows the same authors.

238 *3.2 Laser Ablation Inductively Coupled Plasma Mass Spectrometry*

239 Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS) trace element maps
240 were acquired using a Resonetics M-50-LR 193 nm excimer laser coupled to an Agilent 7700x
241 Quadrupole ICP-MS housed at Adelaide Microscopy, University of Adelaide. Instrument conditions and
242 mapping protocols similar to that employed in this study are outlined in Raimondo et al. (2017). Pre-
243 ablation of each raster scan was completed to minimize the effect of redeposition (19 μm , 75% overlap),
244 followed by 15 s washout and 10 s of background measurement. A beam diameter of 19 μm , line spacing
245 of 19 μm and repetition rate of 10 Hz were employed for sample PU2, resulting in an energy density of
246 3.5 J/cm^2 at the target. Standards were analyzed in duplicate every 2 h during the mapping session,
247 including reference glasses NIST 610 (Pearce et al., 1997; Jochum et al., 2011a) and GSD-1D (Jochum et
248 al., 2011b). A beam diameter of 51 μm was used for all standard analyses, and included 5 pre-ablation

249 shots (51 μm , 75% overlap) followed by 20 s washout, 30 s background measurement and 40 s ablation
250 time. Data acquisition was performed in time-resolved analysis mode as a single continuous experiment.
251 Each analysis comprised a suite of 38 elements, and dwell times were as follows: 0.01 s (Li), 0.002 s (Na,
252 Mg, Al, Si, K, Ca, Mn, Fe, Ni), 0.005 s (Sc, Ti, V, Cr, Nb, Ba, Hf, Th, U), and 0.008 s (Sr, Y, Zr, Ta, Pb, REEs).
253 The total sweep time was 0.297 s. Post-acquisition processing was performed using the software Iolite
254 (Woodhead et al. 2007; Hellstrom et al. 2008; Paton et al. 2011), with data reduction and image
255 processing procedures following those outlined by Raimondo et al. (2017) and Hyppolito et al. (2018).

256 *3.3 Electron Back-Scattered Diffraction and cathodoluminescence mapping*

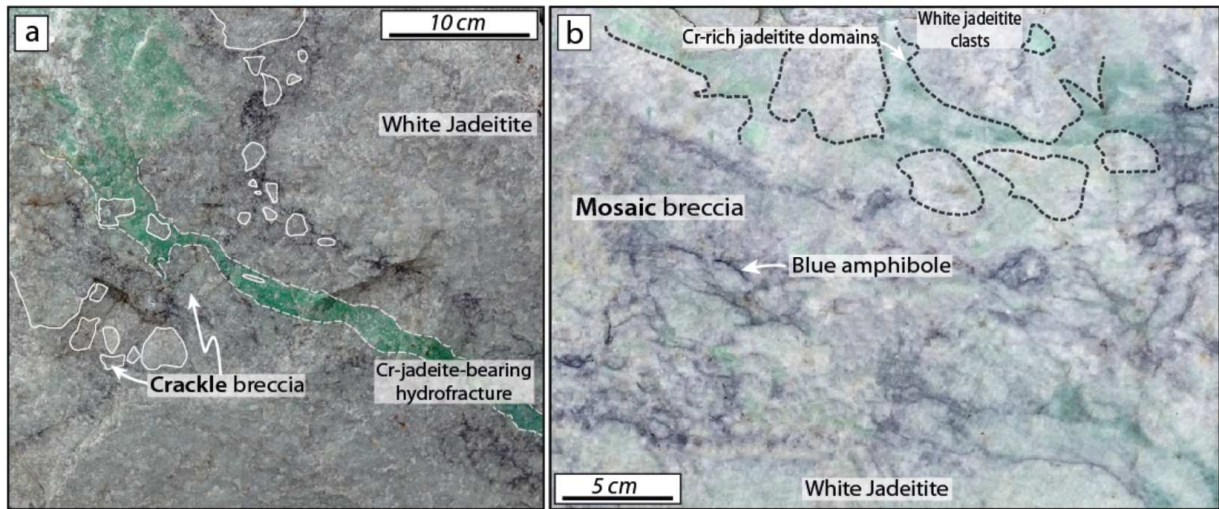
257 Electron Back Scattered Diffraction mapping (EBSD) has been performed at the Laboratoire de
258 Géologie of the Ecole Normale Supérieure of Paris using a ZEISS SIGMA Field Emission Gun Scanning
259 Electron Microscope equipped with an EDX (Energy Dispersive X-ray Spectroscopy; X-MAX) and an EBSD
260 detector (Nordlys Nano, Oxford Instruments). An acceleration voltage of 15 keV, a beam current of 5 nA,
261 an aperture of 120 μm , an inclination of 70°, an acquisition rate of 100 Hz, a working distance of 14 mm
262 and a mapping step size of 2.8 μm were the analytical parameters chosen for the mapping. Data
263 acquisition, post-processing treatment and statistical analysis were performed using Aztec, Channel 5
264 and MTEX software (Bachmann et al., 2010; Bachmann et al., 2011). For noise reduction, every single-
265 pixel isolated data point was removed and followed by denoising MTEX procedures.
266 Cathodoluminescence (CL) mosaic images were acquired using the Cathodyne (NEWTEC) device
267 equipped with a motorized stage, a 12kV and 120 μA plasma, and 2 seconds of image acquisition time.

268 **4. Field constraints on Polar Urals jadeitite body**

269 In the Pus'yerka locality of the Polar Urals, the ore jadeitite body is mostly formed by a white
270 jadeitite core that represents more than 90% of the "dyke" volume (see schematic dyke structure in
271 **Fig.1e**). Locally, remnants of felsic lithologies (comprising an albitite groundmass with paragonite flakes

272 surrounded by a jadeite-bearing corona) were observed, suggesting that the white jadeite formed by
273 replacement of a leucocratic dyke (see the model in Angiboust et al., 2021; see also Dobretsov &
274 Ponomareva, 1968 and Kuznetsov et al., 1986). The whitish jadeite-rich domains comprise weakly to
275 strongly foliated mafic blocks that host millimeter-sized intricate amphibole and phlogopite crystals
276 (sample PU2). These domains, referred hereafter to as dark granofels (sample UR11b), were interpreted
277 by Angiboust et al. (2021) to be produced by the influx of alkali-rich fluids in a warm mantle wedge
278 environment, before the emplacement of the leucocratic dyke. Such amphibole-rich blocks occupying a
279 similar structural position were reported in the Jade Mine Belt area (Myanmar) by Bleek (1908; see also
280 Harlow et al., 2015).

281 The white jadeite is transected by centimeter to decimeter-long cracks filled by phlogopite as well
282 as dark-blueish amphibole-rich domains and emerald-green Cr-rich clinopyroxene (**Fig.2a, b**). While the
283 bulk of the white jadeite mass looks at a first sight microstructurally homogeneous, detailed
284 observations highlight numerous locations where structures are brecciated, ranging from crackle- to
285 mosaic-type breccias (e.g., Woodcock & Mort, 2008). Chaotic breccias, where substantial disruption of
286 the original structure occurred, are also observed in jadeitites and associated dark granofels (**Fig.2b**).
287 Lastly, centimeter- to decimeter-thick, strongly sheared phlogopite-rich metasomatic rinds are observed
288 along the margins of the jadeite body at the contact with the host serpentinite (**Fig.1e**; see also
289 Kuznetsov et al., 1986 and Angiboust et al., 2021). Further south, directly below the Rai-lz peridotite
290 massif, occurs a tectonic *mélange* zone (“Nephrite brook”, Kazak, 1978; **Fig.1a**) that contains blocks of
291 nephrite and rare jadeite within a schistose serpentinite matrix. Sample UR25 represents one of these
292 jadeite pods, considered to be derived from a former “dyke” structure that has been fully disrupted by
293 late subduction zone tectonic deformation.



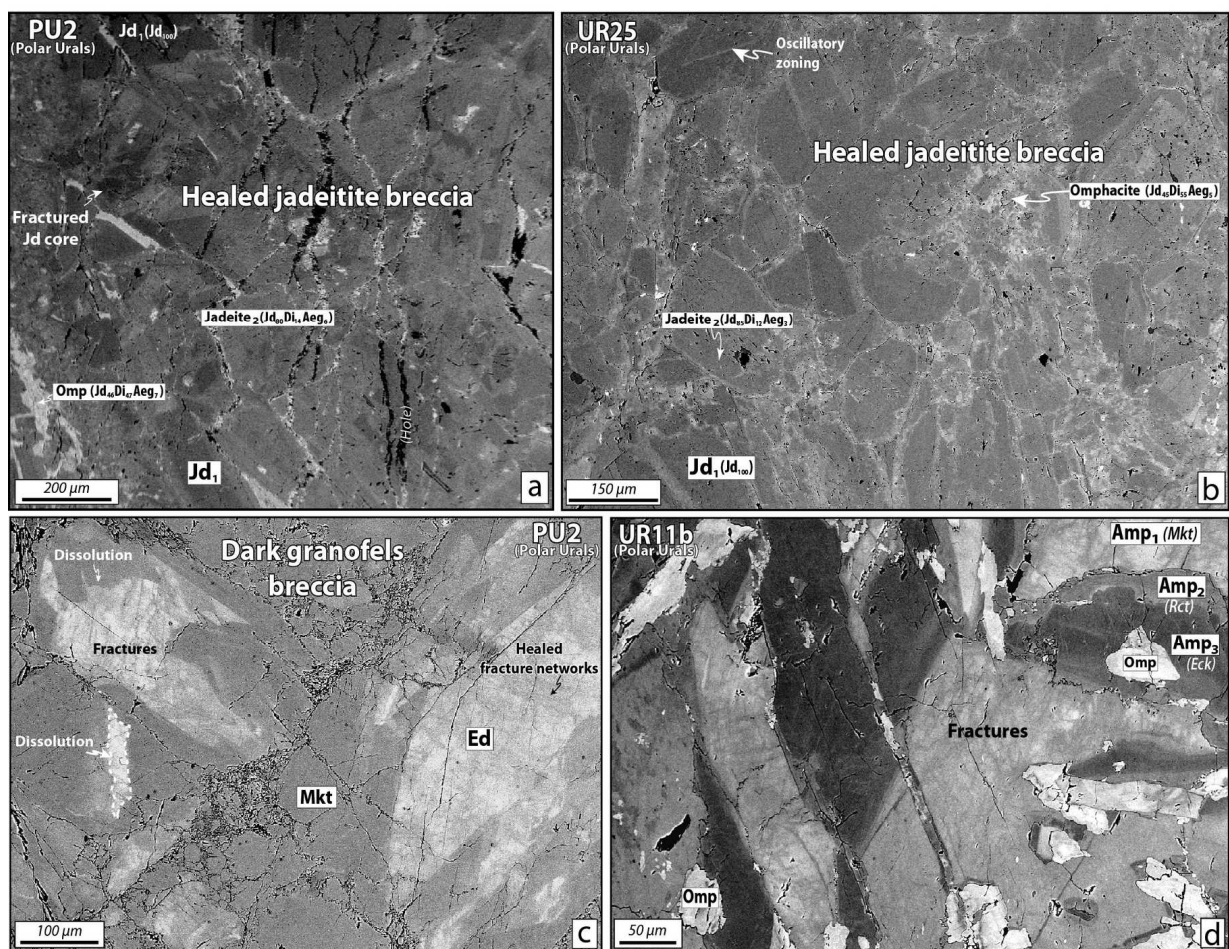
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 295 **Figure 2. a.** Field photo of a brecciated white jadeitite block, showing a discrete crackle breccia texture
 296 transected by a hydrofracture filled with green Cr- rich jadeite. **b.** Field photo highlighting the structure of
 297 a mosaic breccia, with disrupted white jadeitite fragments cemented by a Cr-rich jadeitic clinopyroxene.

298 The nomenclature used in **Table 1** is defined based on the following deformation criteria: Type (I)
 299 corresponds to breccias formed by centimeter- to tens of centimeter-sized clasts exhibiting a substantial
 300 shearing component, fracturing and size reduction through indentation processes. Such breccias can
 301 typically reflect a damage zone deformation pattern. Type (II) brecciated materials are defined as a
 302 highly localized fault zone with evidence of pervasive shearing, grain comminution and flow banding,
 303 with pulverized wall clasts floating in a fine-grained matrix. Such microstructures are generally known in
 304 the vicinity of fault cores and are hereafter termed “foliated cataclasites”. Type (III) corresponds to
 305 hydraulic breccias with millimeter to centimeter-sized clasts (which may have undergone rotation)
 306 cemented by clinopyroxene or amphibole. Space-filling material can be either dendritic, oscillatory, or
 307 strained.

308 5. Structure of Polar Urals jadeitites

309 The bulk of the white jadeitite matrix is formed by idiomorphic, oscillatory and intricate 100–500 μm -
 310 long jadeite crystals (Meng et al., 2011). Their outer rims are commonly lined by Ca-rich, omphacitic

311 compositions as well as interstitial phlogopitic micas (Angiboust et al., 2021; **Fig.S1**; **Table 1** and **Table**
 312 **S1**). Despite an apparently homogeneous macroscopic texture, detailed petrographic investigations
 313 reveal that jadeite crystals exhibit widespread fracturing, dissolution and replacement textures
 314 (**Fig.3a,b**). Vein systems that range in color from white (jadeite composition) to green (Cr-rich jadeite or
 315 Cr-rich omphacite; **Fig.2a**) are ubiquitously found crosscutting the host jadeitite matrix (see also Franz et
 316 al, 2014). These veins are filled by idiomorphic to fibrous clinopyroxene crystals with locally oscillatory
 317 and/or dendritic textures (Kuznetsov et al., 1986; sample UR03b in Angiboust et al., 2021: Type III).
 318 Clinopyroxene crystals from brecciated samples exhibit clear evidence for pressure-solution (as shown by
 319 indentation and truncation textures), as well as solution-precipitation with fracture healing and
 320 overgrowth by a new (generally more omphacitic) clinopyroxene composition (**Fig.3a,b**; **Fig. S2**).



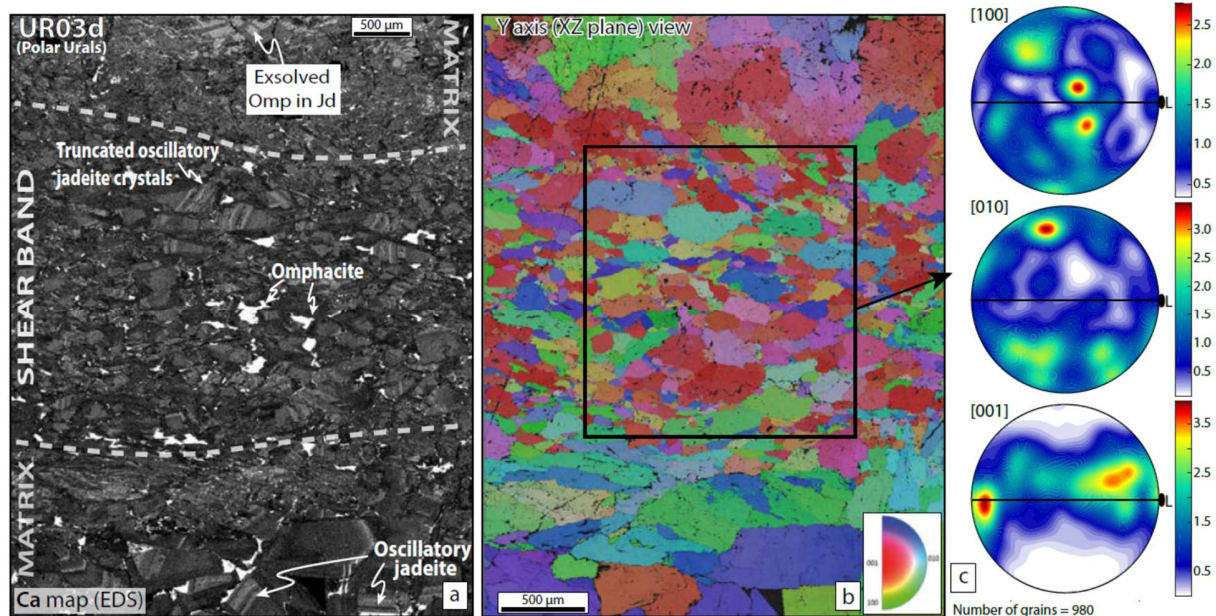
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322 **Figure 3.** Backscattered electron (BSE) images for Polar Urals samples. **a.** White jadeitite specimen (PU2)
323 showing highly disrupted clasts with evidence for fracturing, dissolution, indentation and re-precipitation.
324 Several clinopyroxene generations can be identified with increasing diopside component towards clast
325 rim. **b.** Example of a cryptic jadeitite breccia showing extensive fracturing, comminution and dissolution of
326 dark white jadeitite clasts (locally exhibiting oscillatory zoning pattern). Location of sample UR25 given in
327 figure 1a. **c.** Dark, amphibole-rich granofels that is heavily brecciated, with similar dissolution-
328 reprecipitation features as well as extensive healed microfractures (sample PU2). **d.** Fractures associated
329 with the growth of several amphibole compositions such as richterite (Rct) and eckermannite (Eck)
330 overgrowing magnesio-katophorite (Mkt). Note that omphacite grows lately within a crack in textural
331 equilibrium with Rct and Eck (sample UR11b).

332
333 The studied white jadeitite sample UR03d exhibits coarse-grained, idiomorphic and sector-zoned jadeite
334 crystals in its matrix (**Fig.4a**). These crystals, generally extremely rich in jadeitic molecule (>90 mol.%),
335 are hereafter referred to as Jd_1 . The jadeite growth structures document how early structures have been
336 transected by a localized fine-grained shear zone of oriented jadeite crystals with interstitial omphacite-
337 rich grains. Orientation maps (**Fig.4b**) show that [001] axes are dominant along the fine-grained shear
338 zone, whereas [100] and [010] axes become important approaching the margins of the shear band.
339 Intracrystalline misorientation EBSD maps, depicting misorientation angles between each data point and
340 the mean orientation of the parent crystal (**Fig. S3**), show that grains at the shear zone margins are
341 highly misoriented. In addition, the finer grains within the shear band are consistently devoid of
342 intragranular misorientation except for strongly misoriented larger grains.

343 We interpret that the strongly misoriented clasts may represent fragmented remnants affected by brittle
344 localized shear followed by dynamic recrystallization, resulting in the ubiquitous lack of misorientation in
345 the surrounding finer-grained material. This pattern is highlighted by microscopic observations and
346 strong crystallographic-preferred orientation (CPO), with [001] axes maxima but also forming a weak
347 girdle along the stretching lineation and the foliation plane, respectively, whereas [010] and [100] display

348 maxima subperpendicular to it (normal to and within the foliation plane, respectively; **Fig.4c**). This fabric,
 349 similar SL-type tectonites, is reported in many previous omphacite (a mineral rheologically similar to
 350 jadeite and diopside) fabrics in eclogites (e.g., Philippot & Van Roermund, 1992; Godard & Van
 351 Roermund, 1995; Keppler et al., 2016), compatible with near-plain strain dislocation creep deformation
 352 mechanisms (e.g., Zhang et al., 2006; see also the review paper from Keppler, 2018). In addition,
 353 truncation of oscillatory zoning in jadeite crystals from the shear band denotes the contribution of
 354 solution-precipitation mechanisms. Thus, the fabrics herein observed indicate that a large part of the
 355 material involved in the shear zone (in particular the dark-shaded, fine-grained jadeite rims visible in
 356 **Figure 4a**) grew syn-kinematically, most likely via crystal-plastic deformation processes coupled with
 357 dissolution-precipitation creep.

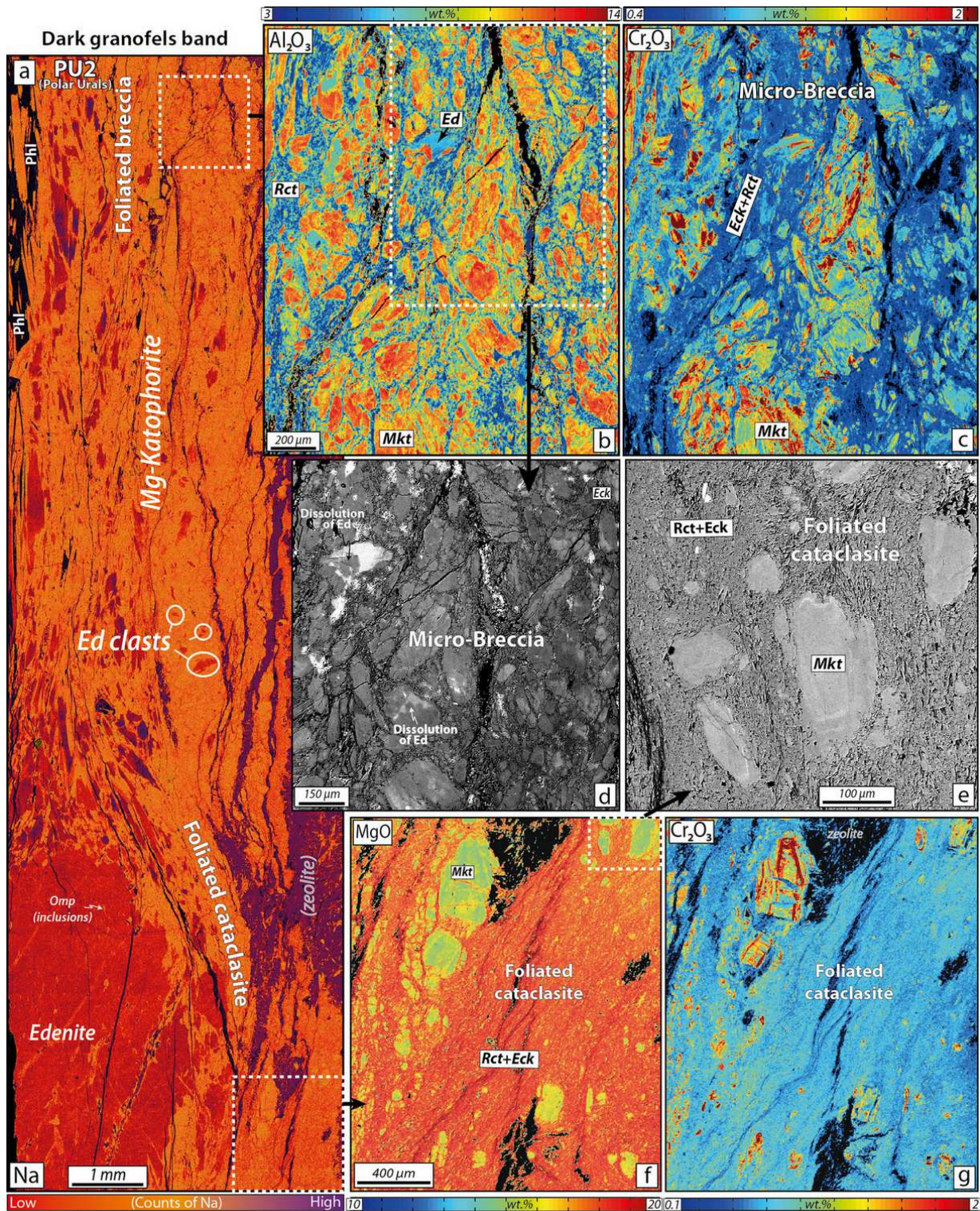


358
 359 **Figure 4. a.** EDS X-ray map (counts of Ca, brighter shades indicating greater elemental concentrations)
 360 showing the structure of a shear band that transects a white jadeite sample (PU3) from the Pus'yerka
 361 deposit. The coarse-grained oscillatory white Jd_1 crystals are deformed within the shear band into
 362 oriented and truncated aggregates of jadeite crystals with local omphacite overgrowths. **b.** EBSD
 363 orientation map colored according to the inverse pole figure key (IPF) of jadeite (bottom right), showing a
 364 shape preferred orientation to the X direction of the strain ellipsoid (i.e., stretching orientation). View (XZ

365 plane) corresponding to the Y axis of the finite strain ellipsoid. c. Pole diagrams of jadeite from the shear
366 band (black square in panel b) represented in an upper hemisphere equal-area projection for [100], [010]
367 and [001] crystallographic axes. Contours are multiples of uniform density distributions. The bold dot on
368 the L axis represents the stretching lineation direction (X direction of the strain ellipsoid), and the black
369 line represents the foliation plane.

370 6. Texture and mineral chemistry of Polar Urals dark granofels

371 The dark granofels found within and along the white jadeitite “dykes” from the Pus’yerka locality exhibits
372 striking evidence for ductile and brittle shearing. Coarse idiomorphic calcic to sodic-calcic amphibole
373 cores of edenite ($\text{NaCa}_2\text{Mg}_5\text{Si}_7\text{AlO}_{22}(\text{OH})_2$) to Mg-katophorite ($\text{Na}(\text{CaNa})\text{Mg}_4\text{AlSi}_7\text{AlO}_{22}(\text{OH})_2$; Amp_1
374 brighter Ca-rich cores in BSE imaging mode) form a dense network of dark, sealed fractures as well as
375 indentation and dissolution-precipitation features (**Fig.3c,d**). Various mutually overgrowing generations
376 of amphiboles ranging from Mg-katophorite (Amp_2 on **Fig.3d**) to sodic amphiboles such as eckermannite
377 ($\text{NaNa}_2(\text{Mg}_4\text{Al})\text{Si}_8\text{O}_{22}(\text{OH})_2$; Amp_3) or richterite ($\text{Na}(\text{CaNa})(\text{Mg,Fe})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$) fill the breccia inter-clast
378 space in apparent textural equilibrium with omphacite (see also Angiboust et al., 2021). In sample PU2, a
379 dark granofels layer is observed (adjacent to a micro-fractured white jadeitite domain), containing
380 oriented sodic-calcic amphibole-phlogopite (\pm omphacite \pm clinocllore) crystals (**Fig.5a**). Pervasive grain
381 size reduction occurred through micro-brecciation (**Fig.5b,c,d**) followed by further comminution that
382 ultimately led to the formation of anastomosing foliated cataclasite networks (**Fig.5e,f,g**; Type II). During
383 (or after) grain size reduction, the large primary amphibole porphyroclasts of edenitic to magnesio-
384 katophoritic compositions (with irregular Cr enrichments; **Fig.5c,g**) were re-equilibrated along their rims
385 and fractures with very fine-grained pulverized domains with richteritic to eckermannitic compositions
386 (**Fig.5c,f**). It thus appears that fracturing and milling occurred after (or during) the formation of Mg-
387 katophorite around the edenitic rims. LA-ICP-MS trace element mapping shows that the Mg-katophorite
388 clasts are enriched in Cr, Nd, Zr and Y, whereas the fine richterite-eckermannite intergrowths are
389 relatively enriched in Li and Ni (see **Fig. S4**).



390
 391 **Figure 5. a.** EDS X-ray map (counts of Na) showing the internal structure of a shear band crosscutting a
 392 dark granofels (sample PU2, Polar Urals). Primary edenite compositions (large clast) are gradually
 393 transformed along their rims into Mg-katophorite to richteritic compositions. Remnants of the original

394 edenite clasts are found floating in the Mg-rich domain, associated with oriented phlogopite (Phl) as well
395 as rare omphacite and clinocllore crystals. **b** and **c**. EPMA X-ray maps showing a close-up on the foliated,
396 matrix-rich, clast-supported chaotic breccia. **d**. BSE image showing dissolution features of the primary
397 edenitic amphibole. **e**. BSE image showing a region of the shear band exhibiting a typical foliated
398 cataclasite texture with Mg-katophorite clasts wrapped in a richterite+eckermannite fine-grained matrix. **f**
399 and **g**. EPMA X-ray maps showing the internal structure of the foliated cataclasite as well as the presence
400 of pre-fracturing Cr sector zoning in Mg-katophorite crystals. Some trace element maps for this sample
401 are provided in **Fig. S4**. Texturally late Na- and Ca-rich zeolites grow as patches and along fractures parallel
402 with the main foliation.

403 An example of Type I mosaic breccia can be seen in **Fig. 6a** (sample UR11b), where clasts of Mg-
404 katophoritic composition were also heavily affected by multiple fracturing events. Remnants from the
405 most pristine amphiboles lie along the edenite-Mg-katophorite transition (see **Table S1** and **Fig. S4** for
406 chemical properties). Omphacitic clinopyroxene is very common as inclusions within clast cores as well
407 as along sealed fractures (**Fig.6d**). Several generations of Mg-katophorite (with the younger exhibiting
408 increasing Na(B) and decreasing Ca contents; **Fig.6c**) can be identified within and around the clasts.
409 Texturally late eckermannite crystals (together with nyboïte: $\text{NaNa}_2(\text{Mg}_3\text{Al}_2)\text{Si}_7\text{AlO}_{22}(\text{OH})_2$) are found
410 filling the clast's cracks as well as replacing quadrangular inclusions in the cores of the fragments (**Fig.6b**;
411 see Angiboust et al., 2021 for details on mineral chemistry). The inter-clast space is filled with fine-
412 grained Mg-katophorite fragments that coexist with zoned clinopyroxene (ranging from jadeite to
413 omphacite in composition; **Fig.6d**) as well as micrometer- to tens of micrometer-sized chromian spinel
414 crystals (surrounded by Cr-clinopyroxene; **Fig.6b**; **Fig. S2**). Electron probe mapping demonstrates that
415 the inter-clast domain is relatively enriched in Al, Na, Ca and Cr with respect to the Mg-katophorite clasts
416 (**Fig. S5**).

417 LA-ICP-MS trace element mapping of the brecciated region from the dark granofels (UR11b) shows
418 Mg-katophorite cores exhibiting Cr oscillations equivalent to sample PU2 (**Fig.5g**), revealing cryptic
419 fractures healed with amphiboles substantially enriched in Zr, Hf, Y, Cr, Li and Ba (**Fig.6e,f,g**; **Fig. S5**).

420 Similar enrichments are distinguished within clinopyroxene-rich fracture-fill material (**Fig. S5**). Patchy
421 enrichment in Li, Ba, Sr, Y and Ce (among other elements) is also visible in the amphibole-clinopyroxene
422 inter-clast domain (see **Fig. S5** for further maps as well as normalized trace-element spider diagrams).

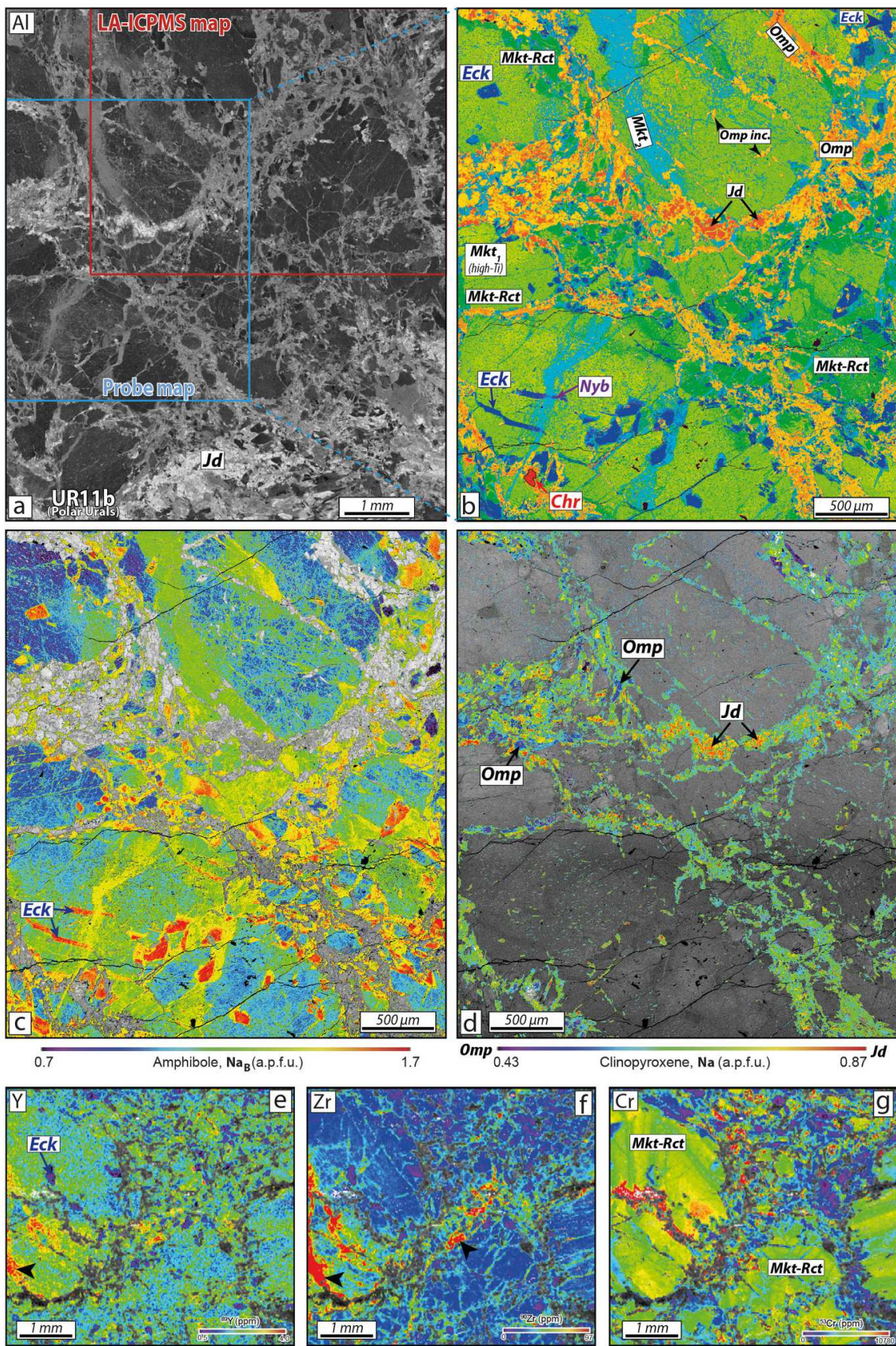
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425

426 **Figure 6. a.** EDS X-ray map (counts of Al, brighter shades indicating greater elemental concentrations)
427 depicting the chaotic breccia texture with the various healed fracture networks, as well as the location of
428 both LA-ICP-MS and electron probe maps (sample UR11b, Polar Urals). **b.** EPMA-based masked phase map
429 showing the various Mg-katophorite generations (green), eckermannite (blue) and nyboïte (violet).
430 Omphacite is in light yellow shades, jadeite in deep orange and chromite (Chr) in red. Note how the
431 various fracture generations are mutually overprinting. **c.** Amphibole masked map showing the Na(B)
432 content. Increasing Na(B) is commonly viewed as reflecting a decreasing P/T ratio (closer to HP-LT
433 conditions). **d.** Masked map of Na content in clinopyroxene showing two distinct compositions in the
434 breccia-filling material. **e, f** and **g.** LA-ICP-MS masked trace element maps (amphiboles only) showing Y, Zr
435 and Cr concentrations (in ppm). The full analytical dataset is given in **Fig. S5**. Black arrows indicate Zr and Y
436 enrichments in the healed fractures and in the inter-clast matrix.

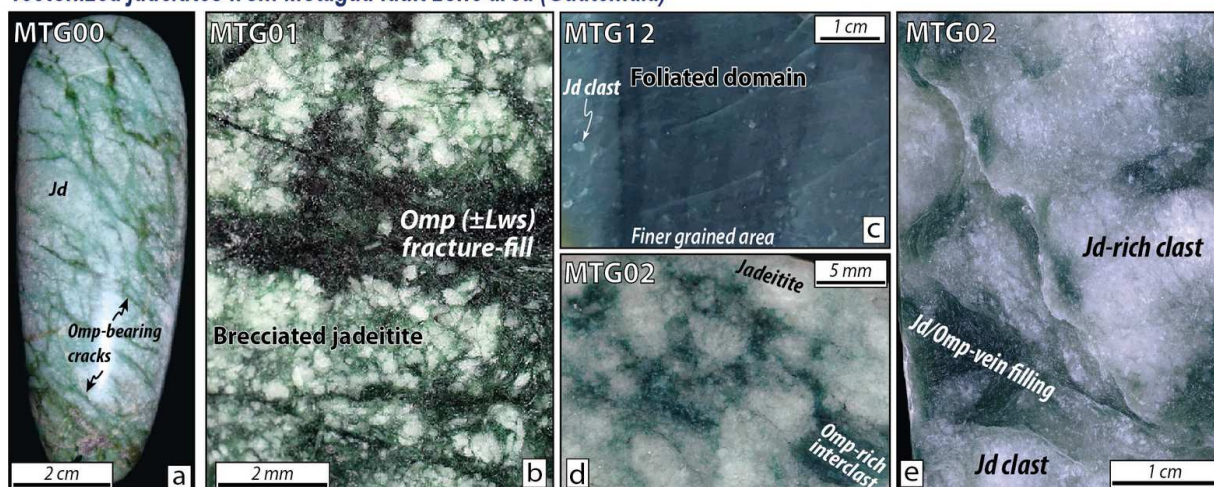
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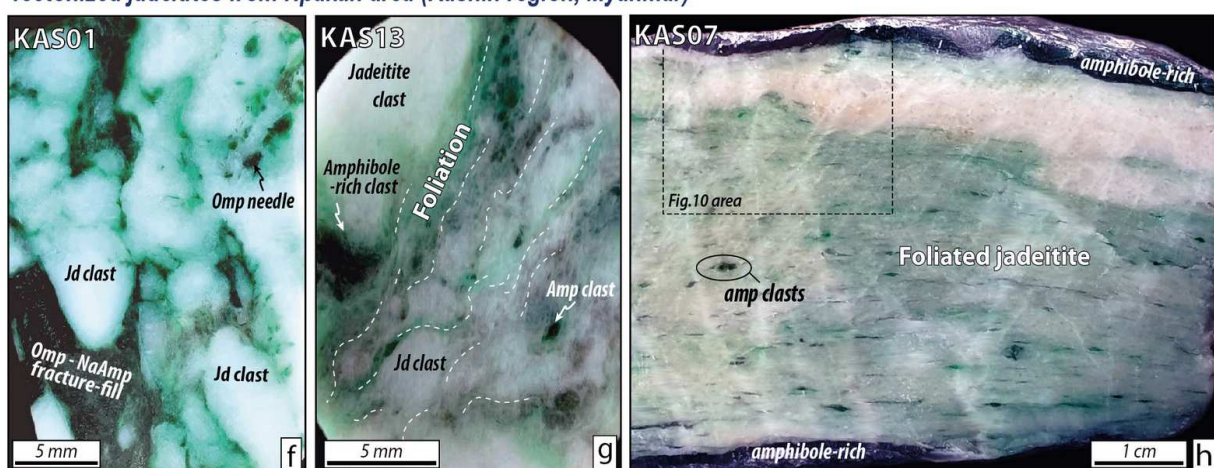
439 **7. Deformation and mineral chemistry of Motagua (Guatemala) jadeitite breccias**

440 Macroscopic evidence for brecciation of the Motagua region jadeitites is widespread in the studied
441 samples (see also Harlow, 2011 and references therein). Three types of deformation patterns were
442 observed: (i) deep green fractures crosscutting white jadeitites (Type I, **Fig.7a,b**); (ii) finely comminuted
443 foliated breccia and cataclastic patterns (Type II, **Fig.7e**); and (iii) hydraulic-like breccias where
444 millimeter- to centimeter-sized angular clasts are infiltrated by a fibrous, oriented jadeitite (bluish)
445 cement (Type III, **Fig.7c,d**; see below). Whereas the clasts forming the breccias are systematically whitish
446 to light-greenish along their rims, the fracture- and breccia-filling material is always darker, ranging from
447 dark green (**Fig.7a,b**) to deep blue (**Fig.7c,d,e**). The inter-clast domain filling these breccias is always
448 enriched in Ca, Mg and Fe and depleted in Na and Al, as a consequence of omphacite growth. The
449 replacement of the original jadeite (Jd_1 , of near pure jadeite composition) by a secondary clinopyroxene
450 (Jd_2 , commonly slightly poorer in jadeite molecule ($80 < Jd_2 < 90$ mol.%) and *Omp*) is pervasive, affecting
451 the entire breccia and not only the clast margins (**Fig.8a,b**). Fracture-filling clinopyroxene exhibits
452 feather-like omphacitic crystals (**Fig. S2**) as well as a texturally strained appearance (**Fig.8a**). The Mg EDS
453 map in **Fig.8a** shows relict white jadeitite clast cores (also visible on the hand specimen image in **Fig.7e**)
454 affected by pervasive re-equilibration (dissolution-precipitation) by Ca-rich jadeite and omphacite
455 compositions along crystal joints, fractures, grain boundaries and micro-faults.

Tectonized jadeitites from Motagua fault zone area (Guatemala)



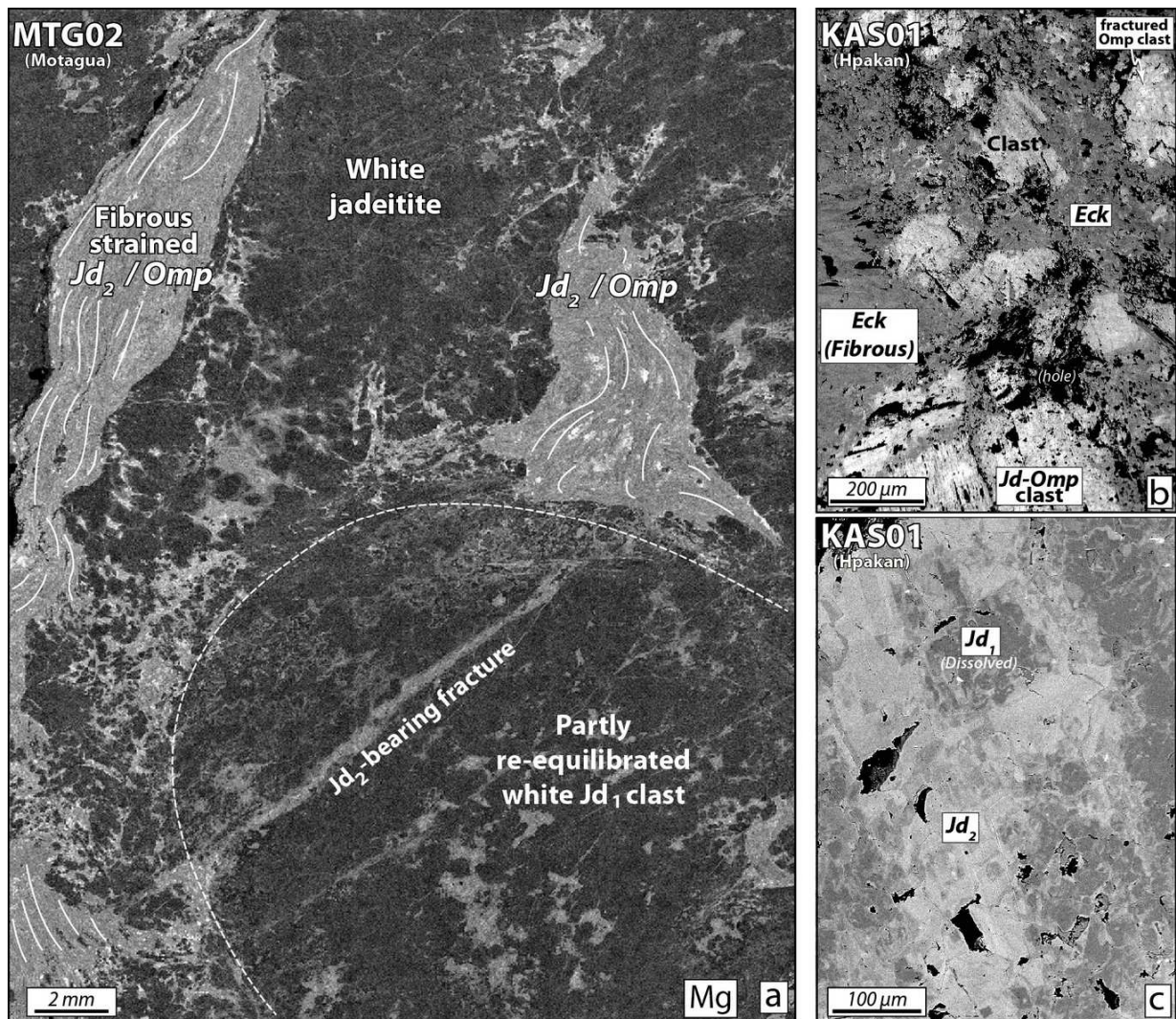
Tectonized jadeitites from Hpakan area (Kashin region, Myanmar)



456
 457 **Figure 7. a.** Polished axe carved in a Motagua fault zone jadeitite. This artefact has been found in El Manati
 458 excavation (Olmec civilization, 2500–500 B.C; Veracruz state, Mexico; length: 105 mm, weight 189.5 g,
 459 density 3.24). It shows numerous omphacite-bearing fractures characteristic of a nearby fault zone system
 460 (Type I). Courtesy of F. Gendron (MNHN, Paris). **b.** Polished rock slab showing numerous white jadeitite
 461 domains rimmed by a dark-green omphacitic clinopyroxene. This sample is cut by a fracture that exhibits
 462 fibrous omphacite crystals as well as rare lawsonite crystals. **c.** Type II jadeitite sample showing the host
 463 that is transected by a foliated domain comprising white jadeitite clasts. **d.** Jadeitite breccia (Type III) with
 464 clasts fractured and healed by a blueish omphacite. **e.** Jadeitite breccia (Type III) with clasts pervasively
 465 fractured and partly recrystallized along their rims by a secondary, greyish jadeite composition, and later
 466 cemented by a deep blue ('Olmec') jadeite composition. **f.** White jadeitite clasts forming a breccia texture
 467 with Cr-enrichments along their rims and cemented by fibrous sodic clinopyroxenes (Type III). **g.**
 468 Moderately foliated breccia showing intermixed dark amphibole-rich clasts and white jadeitite clasts. **h.**

469 Strongly foliated white jadeitite matrix in between two foliated dark granofels bands (Type II). Note the
470 large amount of blue amphibole clasts wrapped in the main clear foliation. The white boudin corresponds
471 to a remnant of the pre-deformation pure jadeitite (Jd_1).

472 In Type II fault rocks, a centimeter-thick layering with variable shades of blue-grey (**Fig.7c**) is
473 observed transecting the original white jadeitite (**Fig.9a**). Pale, clast-like patches (of pure jadeite
474 composition) are macroscopically visible both in the host jadeitite as well as within the deep blue layers
475 (see **Table S1** for representative mineral compositions). Microscopically, the bands are composed of very
476 fine-grained (5–15 μm on average diameter) jadeitite fragments cemented by omphacite-rich
477 compositions in interstitial positions (**Fig.9b,c,d,g**). Interestingly, this apparently pulverized domain also
478 hosts “shard-like” features ranging in size from several tens to hundreds of microns (**Fig. 9e,f,g**). These
479 shards (locally showing ptigmatic folds and contortion) define a weak lamination, resembling the
480 “fiamme” textures known in ignimbrites and other pyroclastic flow deposits. Many of these shards
481 display a fibro-radial intergrowth of jadeite and omphacite fibers that evoke the textures reported in
482 spherulites formed after the devitrification of a former glass (**Fig.9g**). SEM-based surface estimates of
483 these shards yield compositions ranging between those measured for jadeite and omphacite crystals
484 (**Fig.9c,h**). The shard-bearing fault zone domain shown in **Fig.9b** is transected by a dark, very fine-grained
485 omphacitite band that seems connected with the network of omphacite-bearing hydrofractures that
486 transect the rock volume. Remnants from foliated fragments of the host (namely the white jadeitite Jd_1
487 but also the foliated shard-bearing domain) are observed, strongly dissolved, within this omphacitite
488 band. A metasomatic reaction front also appears to have formed at the contact between these two
489 domains (**Fig.9b**).



490

491 **Figure 8. a.** EDS X-ray map (counts of Mg, brighter shades indicating greater elemental concentrations)

492 depicting a Type III hydraulic breccia texture with fibrous clinopyroxene fills (see Fig. S2 for further images;

493 Motagua region, Guatemala). Note how the Jd_1 clasts are corroded and replaced by a secondary jadeite

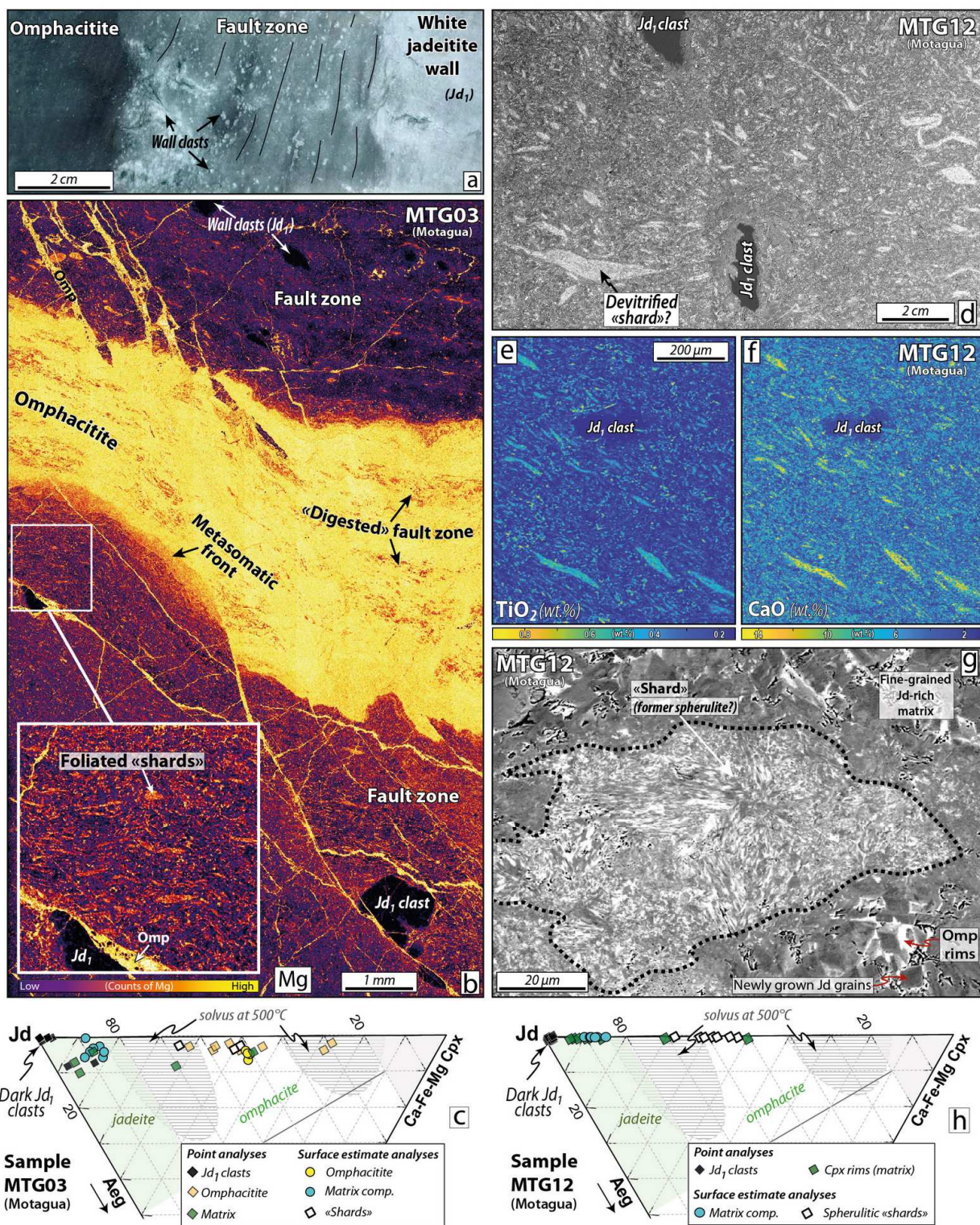
494 (richer in omphacitic component) along grain boundaries, micro-fractures as well as inside the clasts. **b.**

495 BSE image of a brecciated jadeitite with fibrous eckermannite (sodic amphibole) infills (Myanmar). **c.** BSE

496 image showing dissolved primary Jd_1 generation (nearly Jd_{100}) replaced by a secondary Jd_2 composition

497 (slightly more enriched in omphacite).

498



499

500 Figure 9. a. Hand specimen image of a Motagua jadeitite showing deformation features of the pristine
 501 white jadeitite on the right, the foliated clast-bearing domain in the center (Type II texture) and a

502 secondary omphacite band on the left. **b.** EDS X-ray map (counts of Mg) depicting a Type II matrix
503 hosting white jadeitite clasts (in black), foliated “shards” as well as late omphacite-bearing hydrofracture
504 networks. **c.** Clinopyroxene composition in the Jd-Di-Aeg triangle (after Morimoto, 1989; MTG03). Solvus
505 domains for clinopyroxene are derived from Garcia-Casco et al. (2009). **d.** BSE image showing the typical
506 appearance of a Type II foliated cataclasite matrix where shards and partly re-equilibrated white jadeitite
507 clasts coexist. Note how the shards are randomly folded but yet define a mild foliation. **e** and **f.** Quantified
508 EPMA X-ray maps of TiO₂ and CaO contents (in wt.%) showing elemental distributions in a clast and shard-
509 bearing, Type II foliated cataclastic matrix. **g.** High-magnification BSE image showing a shard that exhibits a
510 fibro-radial internal structure that evokes devitrification spherulites. Note how the finely-comminuted
511 clasts forming the pulverized matrix were overgrown by a new (locally faceted) omphacitic clinopyroxene
512 composition during fault consolidation and fluid-enhanced sealing. **h.** Clinopyroxene composition in the
513 Jd-Di-Aeg triangle (sample MTG12). Note how shards are systematically enriched in omphacitic content.

514 **8. Deformation and mineral chemistry of Hpakan (Myanmar) jadeitite breccias**

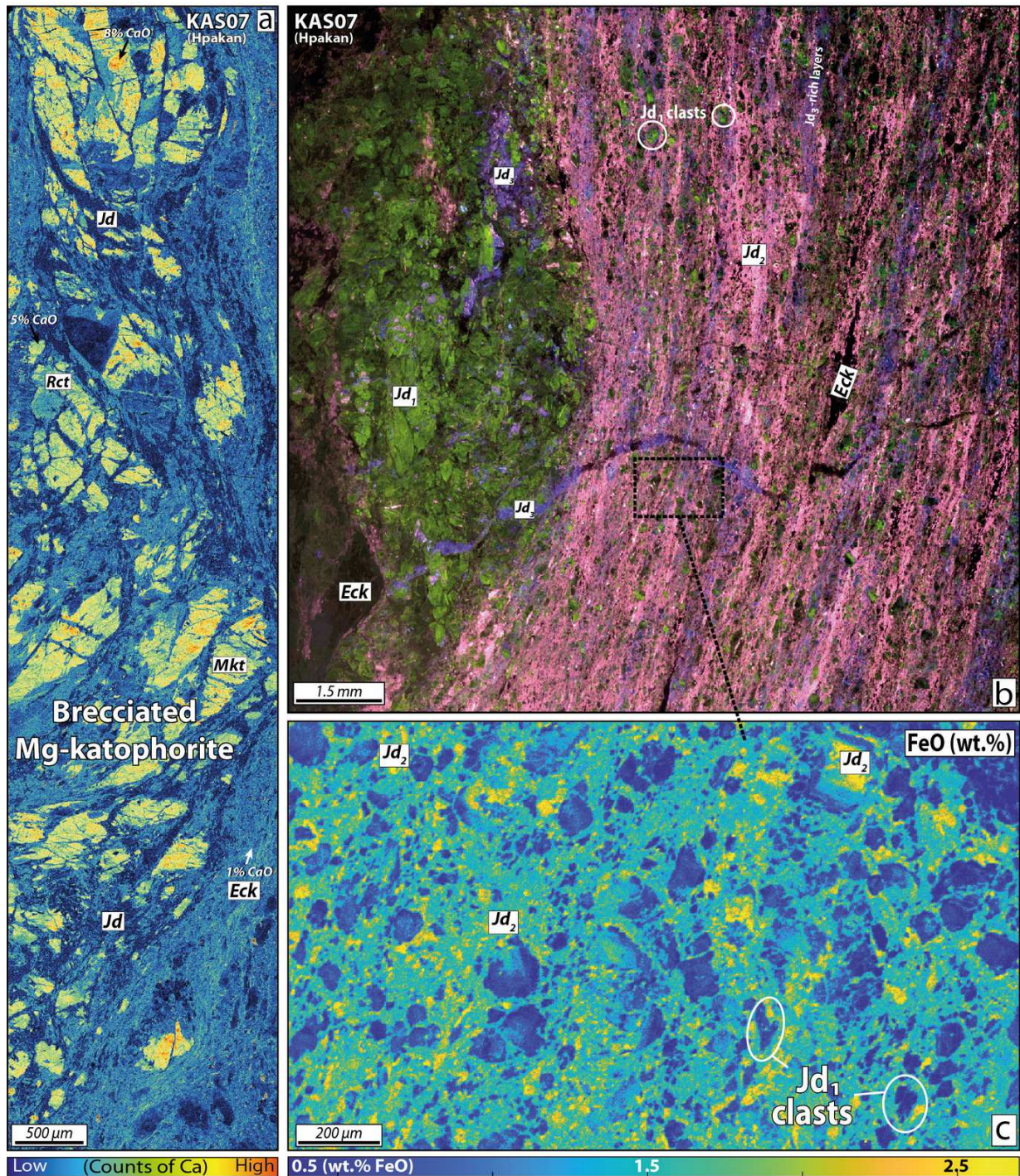
515 Further evidence for brecciation of jadeitites can be obtained observing the hand specimens
516 from Myanmar shown in **Figs 7f, 7g and 7h**. These samples exhibit the most complex sequence of
517 mutually crosscutting brittle events (**Table 1**). In **Fig.7f**, white (and slightly rounded) jadeitite clasts are
518 rimmed by omphacite (light green domains) and wrapped by dark blue amphibole (eckermannite)-
519 clinopyroxene intergrowths (**Fig.8b**). The pristine jadeitic pyroxene composition (Jd_1) is only preserved as
520 islands in the middle of a brighter (in BSE) more omphacitic pyroxene (Jd_2 and Omp; **Fig.8c**). In sample
521 KAS10, dissolved cores exhibiting exsolution features can also be observed, with two pyroxenes forming
522 at the expense of a former one (**Fig. S2**). Jd_1 seems to have been replaced along grain boundaries by a
523 fluid that triggered pervasive re-equilibration (mixed Type I and Type III deformation). In other samples,
524 the same meso-scale structure exhibits a weak, omphacite-rich foliation wrapping remnants of the white
525 jadeitite and dismembered amphibole-rich fragments (**Fig.7g**). This texture suggests that viscous flow
526 may occasionally overprint previously formed brittle (Type I) brecciated zones.

527 Sample KAS07 displays a whitish, foliated band separating two dark granofels domains (**Fig.7h**). The
528 leucocratic band comprises small, flattened dark amphibole fragments oriented parallel with the main

529 foliation (**Fig.10a**), as well as green trails characteristic of Cr-bearing omphacitic clinopyroxenes. A
530 cathodoluminescence image (**Fig.10b**) of the whitish domain reveals the presence of three distinct
531 jadeite-forming events in a Type II pattern deformation zone. The first is a green-colored aggregate (Jd_1 ;
532 white jadeite: $Jd_{91}Quad_5Aeg_4$ on average) wrapped within a pink-colored foliated band (mostly Jd_2) that
533 hosts numerous angular, green-shaded fragments clearly derived from the green Jd_1 aggregate on the
534 left of the image (**Fig.10b**). These green clasts range in size from several tens to hundreds of microns, and
535 display widespread evidence of fracturing, dissolution and re-precipitation/overgrowth by a brighter,
536 slightly more omphacitic composition (see for example **Fig.10c**; $Jd_{84}Quad_6Aeg_{10}$ on average). This fabric
537 evokes a deformation by micro-fracturing, with substantial comminution accompanied with shearing and
538 flow banding of the pink cataclastic domain.

539 In the same sample, an eckermannite-filled extensional vein (Type III) is observed crosscutting at low
540 angles the foliated cataclastic domain (**Fig. S2**). A late pure jadeite formation event (Jd_3 , in purple on
541 **Fig.10b**) also occurs as foliation parallel bands, as patchy domains in the green-colored Jd_1 *augen* on the
542 left of the image, as well as filling late cracks that crosscut all previous features. Sheared amphibole
543 fragments (mostly eckermannite) appear dark in the cathodoluminescence image in the main foliated
544 area (**Fig.10b**). BSE imaging and composition analysis reveal that the amphibole-rich band from the left
545 part of this thin section (**Fig.10a**) locally contains Mg-katophorite cores rimmed by eckermannite or
546 nyböite compositions (see **Table S1** for representative mineral compositions). This brecciated amphibole-
547 rich domain (likely a former Type I) has been cemented by a clinopyroxene-Na-amphibole mixture and
548 foliated parallel to the adjacent foliated cataclastic domain (**Fig.7h, 10b**). A striking feature of this dark
549 domain is the systematic decrease of Ca content in amphibole from the cores (Mg-katophorite; 7 wt.%
550 CaO) to the foliated matrix, where amphiboles are extremely enriched in Na_2O (>11 wt.%) and depleted
551 in CaO (<1.5 wt.%). Similar observations were made in sample KAS06 where cores are edenitic (up to 10

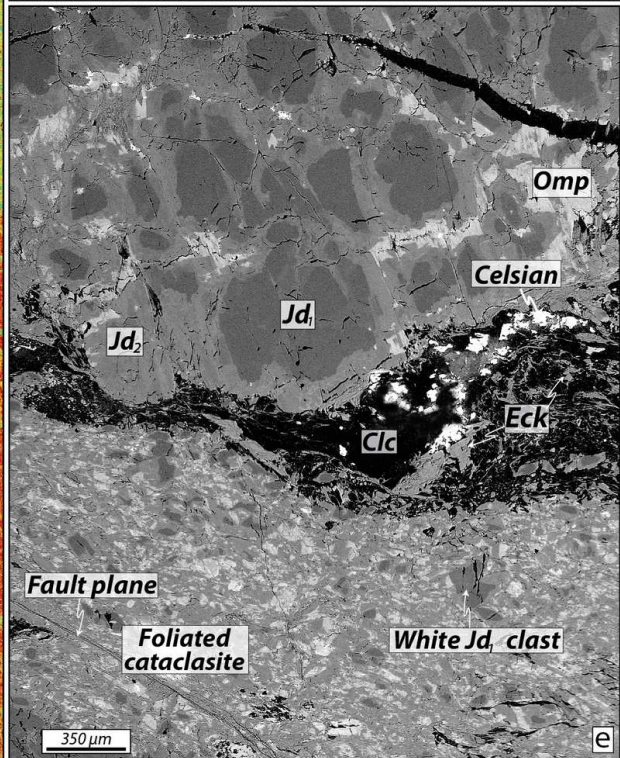
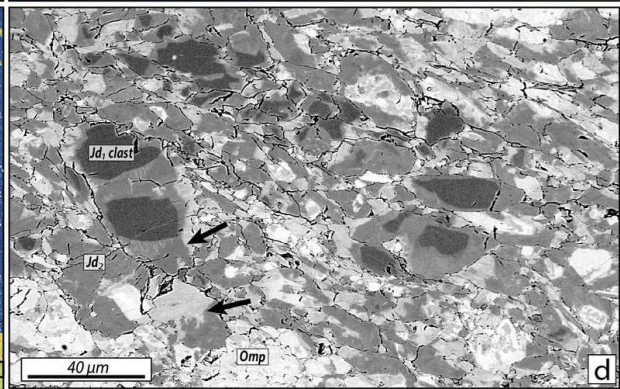
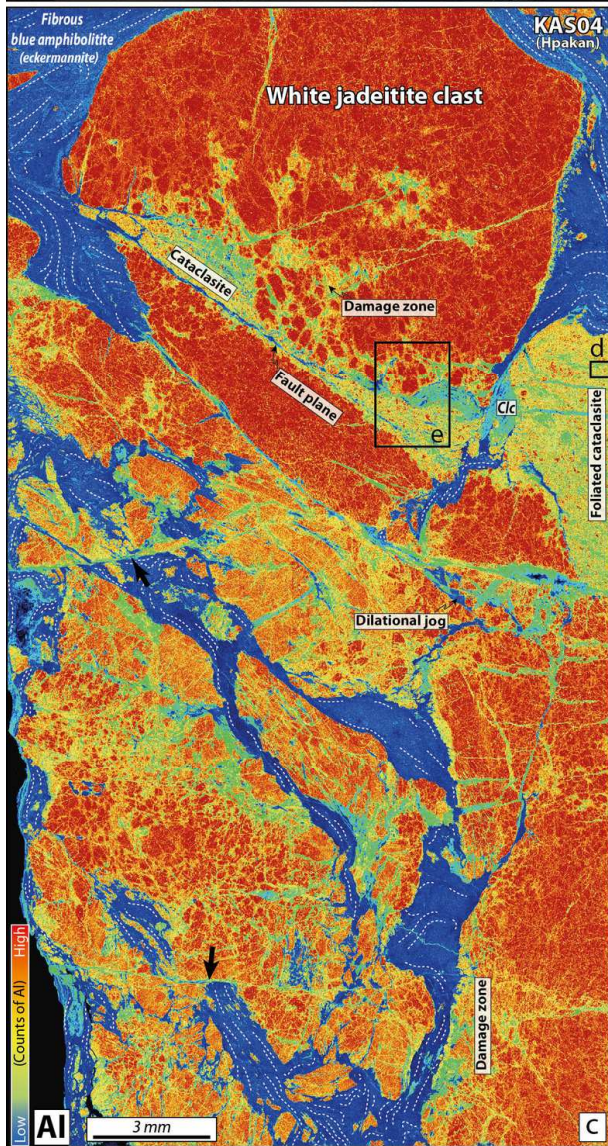
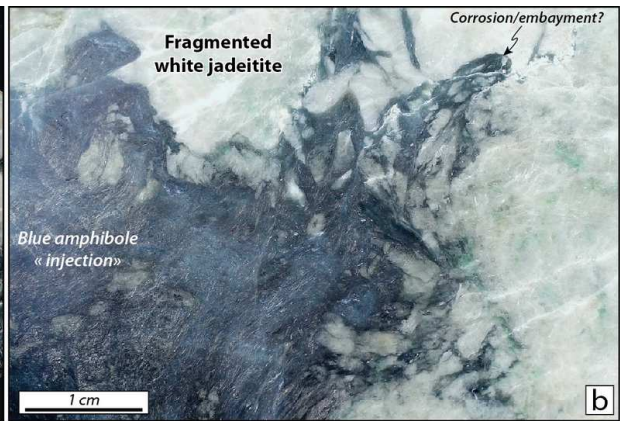
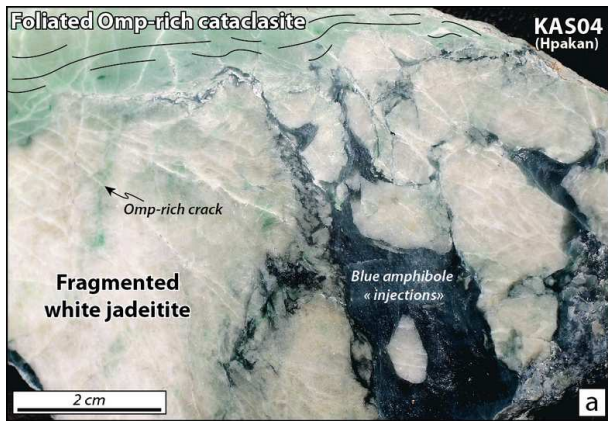
552 wt.% CaO), whereas crystals lining the foliation correspond to eckermannite, also closely associated with
 553 strained jadeite crystals (see Fig. S2 for EDS mapping and further details on mineral chemistry).



554

555 **Figure 10. a.** EDS X-ray map (counts of Ca) showing the internal structure of a foliated and brecciated dark
556 granofels layer with Mg-katophorite cores (locally containing up to 8 wt.% CaO) wrapped within a jadeite-
557 eckermannite-bearing foliation. **b.** Cathodoluminescence image of the whitish, foliated part of sample
558 KAS07. Note the white jadeitite (Jd_1 ; CL-colored in green) fragments now dispersed in the Type II foliated
559 cataclasite matrix (Jd_2 ; CL-colored in pink) and the black trails parallel with the foliation that are mostly
560 made of sodic amphibole (e.g., eckermannite). Late Cr-rich jadeite (Jd_3 ; CL-colored in purple) is also
561 observed parallel with the main foliation as well as along cracks. **c.** EPMA X-ray map showing FeO
562 distribution in the whitish foliated cataclastic domain. Note the random (Fe-poor) Jd_1 clast size distribution
563 as well as the apparent absence of internal deformation (such as plasticity) of Jd_1 clasts.

564 The most compelling evidence pointing to repeated brittle deformation events is exhibited in
565 **Fig.11** (sample KAS04). Three chemically distinct domains can be distinguished: (i) the white jadeitite
566 clasts (likely a former Type I structure); (ii) the blue amphibole fibrous veins that “inject” between
567 breccia clasts (e.g., hydrofracture-filling material); and (iii) a very fine-grained, greenish Type II foliated
568 omphacitite domain at the top of the sample (**Fig.11a, Fig.12f**). Locally, the contact between the blue
569 amphibole domains and the white clasts exhibits structures that evoke Type III hydrofracturing processes
570 (with local embayments pointing to interface-coupled dissolution-precipitation mechanisms; **Fig.11b**;
571 e.g., Putnis & Austrheim, 2013). A large surface of the same sample (KAS04) has been mapped using EDS
572 (**Fig.11c**) in order to image millimeter-scale relationships between these three domains. The largest
573 rounded, Al-rich white jadeite clast (colored in red) is transected along its base by a sharp micro-fault
574 plane, along which the green, very fine-grained material (“foliated cataclasite”) is found. The green
575 omphacitite domain is formed by tens of micrometer-sized crystals (**Fig.11d**) that commonly contain dark
576 clasts (in BSE mode) with a composition similar to the large white jadeitite clasts (**Fig.12f**).
577 Cathodoluminescence imaging reveals that the clasts in the foliated cataclasites have a similar purple CL
578 color as the large white jadeitite clasts, confirming their genetic link (**Fig.12a,b,c**). These clasts are
579 rimmed by two distinct clinopyroxene compositions (Jd_2 and Omp) with increasing omphacitic content
580 (**Fig.11e, 12f**), the interfaces between

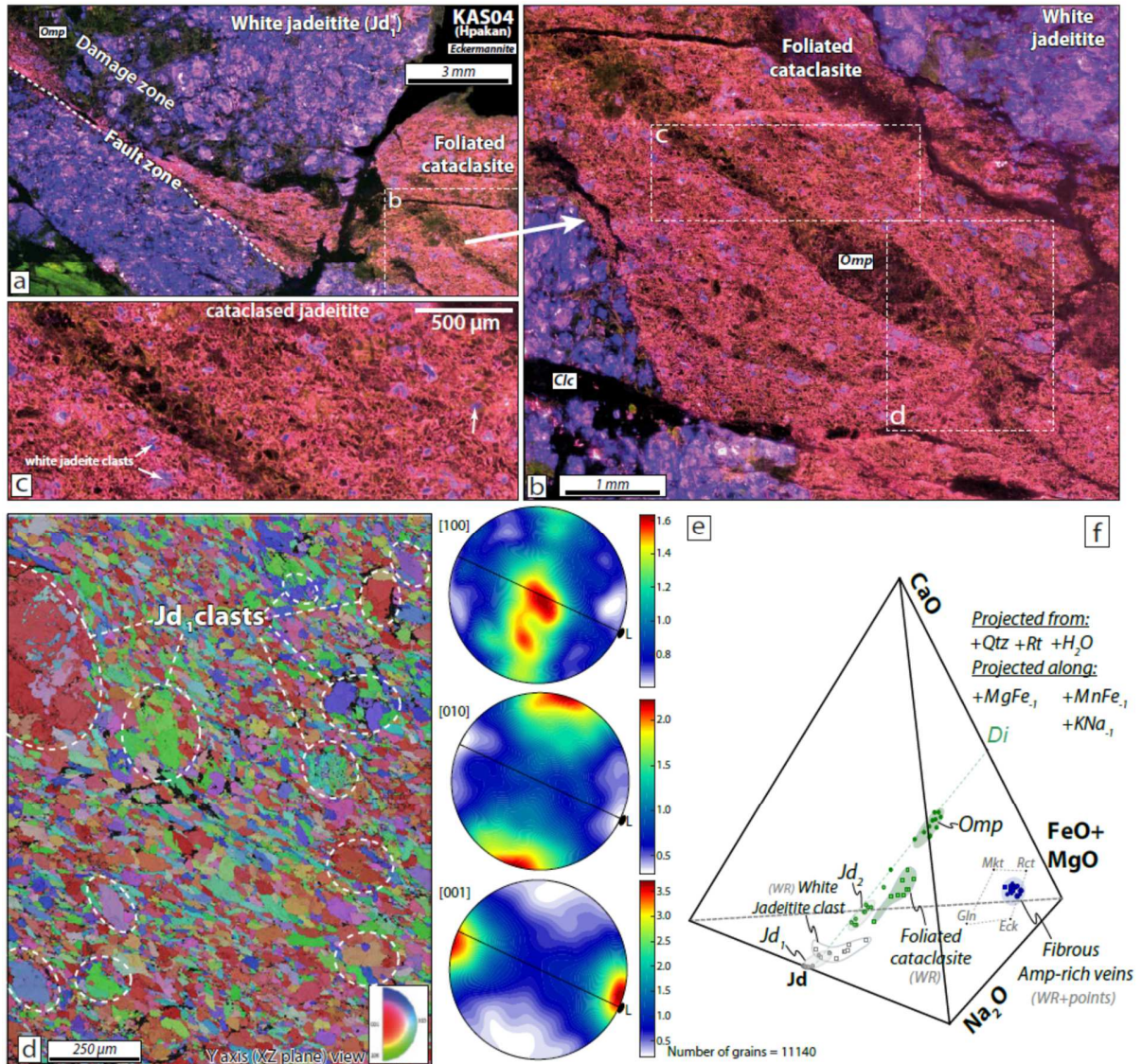


582 **Figure 11. a.** Polished hand specimen of a brecciated jadeitite from Myanmar (KAS04), where white
583 jadeitite clasts are transected by a green, strongly foliated cataclastic domain (Type II). The sample is lately
584 affected by a stage of pervasive, fibrous sodic amphibole (eckermannite) growth, infiltrating along cracks.
585 **b.** Higher magnification image showing how the white jadeitite has been hydrofractured and filled by sodic
586 amphibole, leaving embayments that may be interpreted as resulting from the dissolution by a reactive
587 fluid. Note also the distribution of sodic amphibole fibers infiltrating the Type III breccia. **c.** EDS X-ray map
588 (counts of Al) showing the great complexity of the internal structure in sample KAS04, where several
589 brittle deformation and fluid-rock interaction events can be identified, including deformation features
590 along ancient fault planes with injecting cataclastic material as well as damage zone formation. **d.** BSE
591 image of the foliated cataclasite domain where white Jd_1 clasts are now dissolved and embedded within a
592 Jd_2 -omphacite matrix. **e.** BSE close-up image of the faulted area in panel **c** showing the foliated cataclasite
593 domain, the micro-fault plane as well as the partly re-equilibrated damage zone. The contact between the
594 large white jadeitite clast has been re-activated and filled by a clinocllore-eckermannite-celsian
595 paragenesis.

596 which appear to be dissolved (see black arrows on **Fig.11d**).

597 Elongated patches with omphacitic composition are aligned with the main foliation, and invade the fault
598 as well as its damage zone (**Fig.12a,b**). The foliated nature of this fine-grained jadeite-omphacite rich
599 domain is evidenced using EBSD mapping (**Fig.12d**). This image shows a shape-preferred orientation,
600 with elongated jadeitic clinopyroxene grains (aspect ratios up to 6:1) defining a lineation subparallel to
601 the strong crystallographic-preferred orientation well visible in [001] (**Fig.12e**). Similarly to Polar Urals
602 jadeitites, the [010] crystallographic axes are highly concentrated normal to the foliation plane but [100]
603 are aligned within the foliation plane perpendicular to [010]. This defines a fabric akin to LS-type
604 tectonites, confirming the likely presence of crystal-plastic deformation mechanisms associated with a
605 near-plain strain geometry, as also supported by experimental investigations and numerical simulations
606 (Ulrich & Mainprice, 2005; Zhang et al., 2006). In addition, intragranular misorientation maps show that
607 several highly misoriented grains are surrounded by clusters of finer-grained crystals devoid of internal

608 deformation. The former likely represent remnant fragments, and the latter dynamically recrystallized
 609 grains along the foliation (**Fig.12d, Fig. S3**).



610
 611 **Figure 12. a, b, c.** Cathodoluminescence images showing the relationships between the different textural
 612 domains of sample KAS04. Note how the Jd_1 fragments (pink colors) are dragged within the foliated
 613 cataclasite domain. **d.** EBSD orientation map colored according to the inverse pole figure key (IPF) of
 614 jadeite (bottom right), showing a shape preferred orientation of jadeite to the X direction of the strain
 615 ellipsoid (i.e., stretching orientation). View (XZ plane) corresponding to the Y axis of the finite strain
 616 ellipsoid. **e.** Pole diagrams of jadeite crystals smaller than 100 μm in diameter (larger crystals are
 617 considered as clasts, as indicated in the figure and more detailed in **Fig. S3**), represented in an upper

618 hemisphere equal-area projection for [100], [010] and [001] crystallographic axes. Contours are multiples
619 of uniform density distributions. The bold dot on the L axis represents the stretching lineation direction (X
620 direction of the strain ellipsoid) and the black line the foliation plane. **f.** Projection in a CaO-Al₂O₃-
621 (FeO+MgO)-Na₂O system of the different groups of mineral compositions, as well as local surface
622 estimates of bulk major element composition (WR: whole-rock). Circles are single point measurements
623 and squares are SEM-based surface compositional estimates (WR).

624 A late faulting event that led to the precipitation of blue amphibole as well as clinocllore along
625 micro-fault planes affects the white jadeitite domain and the foliated cataclasites (**Fig.11c,e**). Celsian
626 crystals (a Ba-rich feldspar) are commonly found texturally associated with these apparently late
627 clinocllore-bearing fault zones (black arrows on **Fig.11c**). To summarize, textural relationships visible in
628 sample KAS04 suggest the following deformation sequence: (i) tectonic brecciation of the white jadeitite;
629 (ii) cataclasis of some of the white jadeitite and foliation of this cataclased domain; (iii) re-brecciation
630 (hydraulic) of the volume and precipitation of the Na-amphibole-rich domain between brecciated and
631 cataclased domains; and (iv) crystallization of clinocllore and celsian along late discrete fault zones
632 oblique to the previous fault structures (**Fig.11c**).

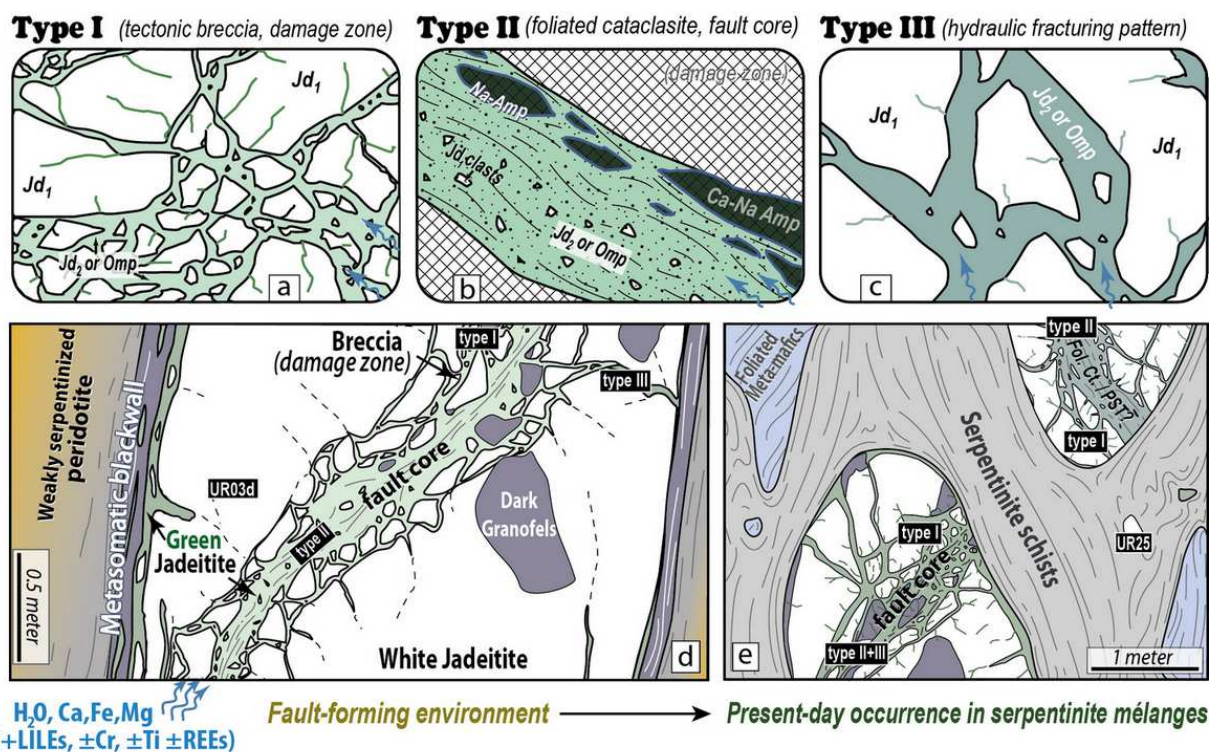
633 9. Discussion

634 9.1 Worldwide jadeitites record mixed deformation regimes

635 Our comparative and analytical investigation reveals that most of the studied samples from Russia,
636 Guatemala and Myanmar display widespread and similar markers of fracturing that have received little
637 attention in previous studies (e.g., Dobretsov & Ponomareva, 1968; Shi et al., 2009b). Brittle deformation
638 markers span a broad range, from Type I breccias typical of damage zones (e.g., **Figs.2, 6, 10a, 13a**), to
639 Type II variably foliated cataclasites that are more diagnostic of strongly localized fault systems (e.g.,
640 Rowe et al., 2011; Angiboust et al., 2015; Oncken et al., in revision; **Figs.5, 9, 13b**), to Type III hydraulic
641 breccias thought to develop at high fluid pressure conditions as extensional veining (e.g. Woodcock et
642 al., 2007). A common feature of all studied materials are arrays of healed mineral fractures commonly

643 coupled with grain size reduction, crystal indentation and dissolution-precipitation processes. This
644 pattern, sometimes hardly visible using optical microscopy, requires the use of specific analytical
645 approaches such as cathodoluminescence imaging or electronic microscopy to be documented (e.g.,
646 **Figs.3, 12**; Shi et al., 2003; Sorensen et al., 2006; Takahashi et al., 2017). Pure extensional fracturing (e.g.,
647 hydrofracturing; Type III; **Fig.13c**) has been identified in several samples, with fragments of the host that
648 appear to have been snatched into clinopyroxene or amphibole-rich veins (**Fig.10b**). When several brittle
649 events can be distinguished, Type III events always occur late in the sequence (**Table 1**). In all localities, it
650 is always the original white jadeitite material (Jd_1) that undergoes shearing-related brecciation and
651 subsequent cementation by a darker clinopyroxene (Jd_2 or *Omp*; **Fig.7e**) and/or by sodic amphiboles
652 (e.g., eckermannite; **Figs.7f, 11a**).

653 Evidence for viscous deformation has been observed in a shear band transecting the coarse-
654 grained Polar Urals white jadeitite body (**Fig.4a**), as well as within Myanmar jadeitites (Shi et al., 2009b).
655 This observation reveals that the white jadeitite body, once crystalized, underwent differential stresses
656 high enough to generate crystal plasticity (most likely via dislocation creep) yielding a CPO. Following the
657 argument developed in Angiboust et al. (2021), it is hypothesized that this shearing occurred at c. 650 °C,
658 i.e., above the antigorite stability field (**Fig.14d**), because no evidence for Cr enrichment was found in
659 this shear band (even though other processes may also contribute to Cr availability in the system; e.g.,
660 Huang et al., 2019). Assuming a strain rate of 10^{-14} s^{-1} (a median value commonly inferred for ductile
661 shear zones; e.g., Pfiffner & Ramsay, 1982) and using the experimentally-based dislocation creep flow
662 laws from Orzol et al. (2006) and Zhang et al. (2006), it is possible to estimate the differential stresses
663 required for creating the shear band of sample UR03 to be between 125 and 40 MPa, respectively. Note
664 that these values likely represent an upper bound since other competing deformation mechanisms such
665 as pressure solution creep would likely be activated at lower stress values (e.g. Godard & Van Roermund,
666 1995). The implications of these values will be discussed hereafter.



667
 668 **Figure 13.** a, b and c. Sketches showing the key features of the three main categories of tectonized
 669 jadeitites identified in this comparative study. d. Sketch depicting the white jadeitite structure (derived
 670 from observations from the Pus'yerka body) at the time of brecciation, indicating the various elements
 671 identified in the field and their position with respect to the main fault and host ultramafics. e. Sketch
 672 highlighting the structure of jadeitite blocks as exposed in HP-LT serpentinite mélanges, where they are
 673 commonly associated with blueschist and/or eclogite-facies crustal rocks. We posit that the faulting
 674 frequently observed within jadeitite blocks has been acquired earlier, when the jadeitite was a dyke-like
 675 structure hosted by weakly serpentinized mantle wedge peridotites.

676
 677 The CPO observed in KAS04 sample and other Type II samples likely results from a combination
 678 of dislocation creep and solution-precipitation processes in a dominant simple shear regime but also
 679 with some degree of pure shear, as suggested by the observed LS-type fabric but also a weak girdle of
 680 [001], respectively (**Figs.4c,12e**). These deformation mechanisms have been already documented in
 681 eclogite-facies clinopyroxene by Godard & Van Roermund (1995; see also Ulrich and Mainprice, 2005)
 682 (e.g., **Fig.12b**). It is clear from the fragmented jadeitite clast distribution (along with the

683 cathodoluminescence color of the various crystals; **Fig.12b**) that cataclastic flow (during brittle
684 deformation) operated through localized faulting *before* CPO formation (see also Oncken et al., in
685 revision; **Fig.10c**). In other words, we infer that the observed foliated cataclasite structure is the
686 protracted result of consecutive brittle flow followed by slower, viscous creep. This sequence of events
687 may be the record of one or several seismic cycles where fast co-seismic slip is immediately followed by
688 a slower after-slip event (e.g., Sibson, 1986; Rowe et al., 2011). Indeed, many of the reported fabrics
689 resemble the structures documented in fault cores at shallower/colder environments where multiple
690 episodes of particle fluidization are reported on the principal slip zone (**Figs.5f, 10c**; e.g., Sibson, 1986;
691 Snoke et al., 2014; Cox & Munroe, 2016; Muñoz-Montecinos et al., 2021a). Interestingly, the main
692 brecciation/cataclasis event in sample KAS04 is followed by an episode of hydraulic brecciation where
693 Na-amphibole precipitates Type III; **Fig. S2**). Similar deformation patterns have been observed in dark
694 granofels that also comprise breccias (**Fig.6**), foliated cataclasites (**Fig.5**) and multiple fracturing events
695 (**Fig.3c**; **Fig. S5**; **Table 1**).

696 Exceptionally, the finding of elongated spherulitic shards in a fine-grained sheared domain
697 (**Figs.7c,9d**) – possibly formed by the devitrification of former glass fragments – may be indicative of
698 former pseudotachylytes (i.e., a glassy injection domain formed by melting during fault slip at seismic
699 strain rates; e.g., Austrheim & Boundy, 1994; Austrheim & Andersen, 2004). The *Jd1* clasts found floating
700 in the comminuted matrix display extensive fracturing, grain-size reduction and stress-driven corrosion
701 that characterize pseudotachylyte-bearing systems (e.g., Kirkpatrick & Rowe, 2013). Glassy shards similar
702 to those shown in **Fig.9g** are mostly reported in ignimbrite deposits (e.g., Bull & McPhie, 2007) and in
703 rocks from ancient impact craters (e.g., Dressler & Reimold, 2001). Although, to the best of our
704 knowledge, such shard-like textures have not been described in ancient fault zones, the discovery of
705 unreported textures must be expected because the formation environment of these slip events strongly
706 differs from other paleo-earthquake reports made in the downgoing lithosphere (see below).

707 *9.2 Jadeitite formation and pressure-temperature conditions of brittle deformation*

708 In most serpentinite mélangé settings, the original structure at the time of jadeitite formation has
709 been obliterated or disrupted by subduction zone tectonics (e.g., Harlow et al., 2015). Whereas in several
710 cases jadeitites appear to be directly precipitated from hydrothermal fluids (“P-type” jadeitite, according
711 the classification from Tsujimori & Harlow, 2012; e.g., Garcia-Casco et al., 2009; Harlow et al., 2011),
712 some localities preserve textural features typical of a replacement process (“R-type” jadeitite according
713 to the same classification), including igneous zircon crystals with older crystallization ages (Yui et al.,
714 2012; Hertwig et al., 2016) or the presence of incompletely “jadeitized” trondhjemitic remnants in the
715 dyke volume (Angiboust et al., 2021). Indeed, early observations from Dobretsov & Ponomareva (1968)
716 and Kuznetsov et al. (1986) in Russia, as well as Bleek (1908) and Chhibber (1934) in Myanmar, reveal
717 that original jadeitite may form as dyke-like structures. It is plausible that these dykes were jadeitized
718 during long-term cooling within the high-pressure field, coeval with replacement of a former felsic
719 intrusive body formed earlier in the mantle wedge above the subduction interface, and possibly shortly
720 after subduction initiation in a warm thermal regime (see the recent model by Angiboust et al., 2021; see
721 also Meng et al., 2011 and **Fig.14**). Such secular cooling has also been documented above the
722 metamorphic soles from New Caledonia (e.g. Soret et al., 2016) and Oman ophiolites (e.g. Prigent et al.,
723 2018).

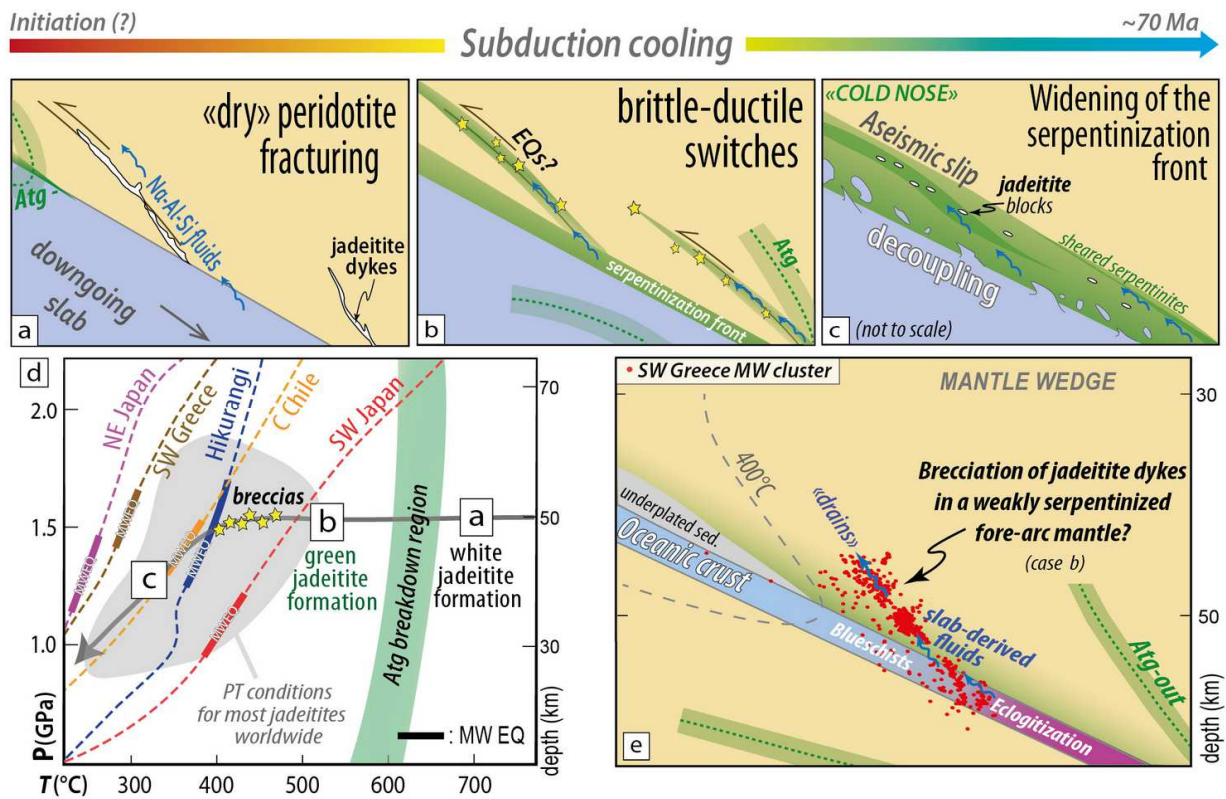
724 A number of recent studies (e.g., Garcia-Casco et al., 2009; Schertl et al., 2012; Angiboust et al., 2021)
725 have suggested that jadeitites may commonly form at temperatures far above 500 °C, perhaps up to 700
726 °C. Garcia-Casco et al. (2009) and Cárdenas-Párraga et al. (2012) have also suggested that exsolution
727 features, as observed in sample KAS10 (**Fig. S2**), could evidence the cooling of a clinopyroxene of
728 intermediate jadeite-omphacite composition below the solvus at 500 °C. This implies that white
729 jadeitites including the Myanmar samples formed at temperatures higher than generally expected.
730 Consistent with this interpretation, Harlow & Sorensen (2005) suggested that the whitish color of the

731 relatively high temperature jadeitite may be due to Cr stability in the host ultramafic chromite and
732 clinopyroxene crystals. In other words, it is likely that Cr-poor white jadeitite bodies formed at T
733 conditions above the antigorite stability field, i.e., before serpentinization of the host ultramafic (**Fig.14**),
734 even though some Cr may also be hosted in high temperature chloritite blackwall (not documented in
735 the studied localities; see Cardenas-Parraga et al., 2012). Following this logic, it can be concluded that
736 the systematic observation of darker-shaded clinopyroxene precipitates/overgrowths within breccias or
737 cataclasites in many jadeitite localities indicate that fracturing and fluid-rock interaction events occurred
738 during long-term cooling at $T < 600$ °C (e.g., Garcia-Casco et al., 2009; Schertl et al., 2012; this study).

739 Whereas breccias and cataclasites are classically reported for rather shallow – and hence colder –
740 environments (<15 km depth, $T < 300$ °C; Sibson, 1986), a number of field examples have recently shown
741 that such features can form at depths greater than 30 km in the downgoing subducted crust (e.g.,
742 Angiboust et al., 2012; 2015; Hertgen et al., 2017; Muñoz-Montecinos et al., 2021a; Oncken et al., in
743 revision). Minerals filling the studied breccias and growing within these fault zones encompass (i)
744 omphacite (and locally lawsonite) in Guatemalan samples; (ii) omphacite, jadeite, phlogopite and Na-rich
745 amphiboles in the Polar Urals samples; and (iii) Na-amphiboles, jadeitic and omphacitic clinopyroxenes in
746 Myanmar samples. These parageneses point to HP-LT conditions ranging between 1–2 GPa and 300–500
747 °C as documented in the reviews of Tsujimori & Harlow (2012) and Harlow et al. (2015; **Fig.14d**).
748 Angiboust et al. (2021) have demonstrated that the phlogopite-bearing fracturing event seen
749 crosscutting Polar Urals jadeitites occurred at least 15 Ma after the HT jadeite-forming event (see also
750 Meng et al., 2011), thus providing an independent confirmation of the secondary nature of the
751 brecciation process. We conclude that in the studied set of samples – and perhaps as a rule in fossil and
752 active subduction systems – brittle deformation of jadeitites is a deep process occurring at or near peak-
753 burial depths in a cooling environment (as suggested in **Fig.14**), rather than as a consequence of late
754 exhumation-related deformation. In order to undergo brittle deformation, the rock must be in-situ and

755 surrounded by mechanically strong material, rather than as a tectonic block floating in a serpentinite-
 756 matrix mélangé where stress is dissipated in the weak matrix.

757



758

759 **Figure 14. a, b and c.** Sketches depicting the long-term tectonic evolution of jadeite “dykes” along the
 760 base of the mantle wedge, from their formation (by replacement of a previous dyke structure or from
 761 direct precipitation from a fluid) to their fracturing, dismembering and wrapping as blocks in a chaotic
 762 serpentinite mélangé. Previous geochronological data (Angiboust et al., 2021) confirms that several tens
 763 of millions of years of subduction activity and mantle wedge hydration are required to achieve a structure
 764 as shown in panel (c). Atg-: antigorite dehydration reaction. **d.** Pressure-Temperature diagram indicating
 765 the conditions of formation of the main white jadeite bodies documented worldwide, and the conditions
 766 of the green jadeite tectonic overprint (and associated sodic amphibole generation). After, and only after
 767 brecciation (yellow stars), the dyke fragments can be exhumed along the plate interface, provided that
 768 the hangingwall serpentinization rate is sufficiently high to enable return-flow of these dense blocks in a
 769 buoyant, serpentinitized matrix (e.g., Gerya et al., 2002; Guillot et al., 2009). P-T conditions for jadeitites

770 are sourced from the compilations by Tsujimori & Harlow (2012) and Harlow et al. (2015). Also shown are
771 the slab-top temperature estimates as calculated by Antriasian et al. (2019, Hikurangi), Wang et al. (2020,
772 central Chile), Halpaap et al. (2019, W Greece) and Peacock (2003, NE and SW Japan). References used for
773 the depth of mantle wedge earthquakes (MW EQ) are provided in text. **e.** Location of one of the MW
774 clusters identified by Halpaap et al. (2019) in W. Greece and interpreted as marking the trace of plate-
775 interface fluids migrating along a vent structure. The formation of this EQ nest (corresponding to the
776 yellow stars from panel d) likely implies that the serpentinization ratio of the subduction interface hanging
777 wall is very low, thus pointing to a structure analogous to the sketch drawn in panel (b).

778 *9.3 Origin of brecciated dark granofels and relationships with fluid-rock interaction events*

779 A striking similarity characterizes the textural relationships between the dark granofels and
780 jadeitites from the Polar Urals and Myanmar. In both localities, they occur as thin layers or pods of
781 “amphibolites” of several tens of centimeters (Chhibber, 1934; Angiboust et al., 2021; **Fig.1e**), generally
782 striking parallel to the main “dyke” foliation and deformed coeval with the associated jadeitites. Their
783 amphibole compositional patterns are also nearly identical, with relicts of Ca-rich amphiboles in the
784 granofels mineral cores (mostly edenite and Mg-katophorite), and Ca-poor, Na-rich amphiboles along
785 their rims and along highly metasomatized domains (Shi et al., 2003; **Fig. S2**). Field observations
786 (Chhibber, 1934; Angiboust et al., 2021) and petrological investigations suggest that two distinct dark
787 granofels occurrences may exist: (i) as disaggregated fragments (the “schistose amphibolite inclusions”
788 from Chhibber, 1934; see also Harlow et al., 2015) inherited from a pre-jadeitization event; and (ii) as
789 amphibole-rich blackwalls formed at the contact between the jadeitite body and the host ultramafic (Shi
790 et al., 2003). The latter should in theory have formed lately at $T < 350$ °C (Shi et al., 2003) and only exhibit
791 Na-amphiboles (since no Ca chemical potential gradient exists between ultramafics and jadeitites), thus
792 hampering the formation of calcic and sodic-calcic amphiboles such as edenite or Mg-katophorite in the
793 reaction blackwall. The former should instead contain these sodic-calcic amphibole species (e.g.,
794 **Figs.5a,6b,10a**) as remnants of an early HT event, associated with the infiltration of alkali-rich fluids
795 and/or melts (likely trondhjemitic in composition) within ultramafics in a subduction initiation setting.

796 This process is known to occur in anomalously high subduction thermal regimes (see also the discussion
797 in Angiboust et al., 2021 and similar structures and rock assemblages in Lazaro et al., 2011 and Soret et
798 al., 2016). Brecciation, followed by boudinage, late tectonic disaggregation and block-in-matrix
799 deformation, most likely blurred the primary contacts, hampering an accurate reconstruction of the
800 geometry of the original jadeitite-dark granofels boundaries (e.g., **Fig.10a**).

801 *9.4 Fluid ingress and metasomatic imprint*

802 A common feature reported in jadeitites worldwide is the enrichment in Ca, Fe and Mg of jadeitic
803 clinopyroxenes due to the infiltration of Ca-Fe-Mg-rich fluids associated with dissolution-precipitation
804 processes during late fluid-rock interaction events (e.g., Sorensen et al., 2006; Garcia-Casco et al., 2009;
805 Harlow et al., 2011; Cardenas-Parraga et al., 2012; Angiboust et al., 2021; this study). The formation of
806 dissolution features (**Fig.3c**), re-equilibration of clast rims (**Fig.8a**) and the metasomatic overprint left in
807 the rock record (**Fig.9b**) are witnesses to the apparently corrosive nature of the infiltrating fluids, which
808 were at thermodynamic disequilibrium with respect to the infiltrated material. The absence of
809 carbonates in most jadeitite occurrences suggests that the fluids were CO₂-poor, most likely because the
810 CO₂ precipitated as carbonates before reaching the jadeitite-forming location. Substantial Cr
811 enrichments are also visible in some of the sheared domains and between breccia fragments (as shown
812 for instance by the growth of chromian spinel around Mg-katophorite clasts in sample UR11b; **Fig.6b**; see
813 also the green overgrowths around white jadeitite clast in **Fig.7f**).

814 As discussed earlier, the release of Cr may be related to serpentinization of the host peridotite
815 during fluid-rock interaction. It is important to note that (i) jadeite/omphacite is found growing within
816 brecciated white jadeitites (**Figs.3b, 8a and 11c**) and amphibole-rich dark granofels (**Fig.6d**); and (ii)
817 amphibole growth is observed in brecciated dark granofels (**Fig.5b**) and between white jadeitite breccia
818 fragments (**Figs.7f,11a,11b**). From these observations, it appears that the incoming fluid composition

819 was not buffered by the wall-rock composition and that brittle events contributed to a chemical and
820 mechanical homogenization of the original dyke-forming lithologies. Furthermore, the widespread
821 formation of texturally late micas (e.g., phlogopite in the Polar Urals: Angiboust et al., 2021, see also **Fig.**
822 **S1**; phengite or phlogopite in Guatemala: Harlow et al., 2011; Flores et al., 2013; white mica in W. Japan:
823 Shigeno et al., 2012) associated with enrichments in LILEs and other fluid mobile elements (**Figs. S3 and**
824 **S4**) confirm that fluid sources were variable in time, switching from a mafic oceanic crust signature to a
825 more hybridized composition with transient highs in sedimentary input (e.g., Sorensen et al., 2006, 2010;
826 Morishita et al., 2007; Meng et al., 2011, 2016; Harlow et al., 2015; Chen et al., 2018; Cardenas-Parraga
827 et al., 2012). This temporal variability of incoming fluid chemistry from a more mafic to a more
828 sedimentary signature could likely be explained by variations in the amount of sedimentary material
829 undergoing devolatilization reactions in the underlying subduction channel (e.g., Bebout, 2007;
830 Scambelluri et al., 2019). This chronological sequence confirms that the same fluid pathways (namely the
831 jadeitite “dykes”) were used over several millions of years as major drains for highly pressurized plate-
832 interface fluids, on their way to the partly hydrated mantle wedge (e.g. Doglioni et al., 2009; Spandler &
833 Pirard, 2013).

834 The distribution of jadeitite “dykes” as seen in the Polar Urals (**Fig.1b**; see also Angiboust et al.,
835 2021) demonstrates that fluid pathways are rather discrete in the overlying plate (i.e., jadeitite does not
836 precipitate randomly everywhere in the fore-arc mantle). It also indicates that less mechanical energy is
837 required to re-fracture a pre-existing physical discontinuity (e.g., a dyke-like structure) than creating new
838 channels for draining plate-interface fluids. This implicitly requires that breccia sealing must have been
839 faster than rupture recurrence to explain the mutual crosscutting relationships documented between
840 the various events identified in our study (see also Woodcock et al., 2007). It seems clear that nearly
841 lithostatic pore fluid pressure was sustained throughout the entire jadeitite drain activity as
842 demonstrated by the ubiquitous evidence for fluid-rock interaction in the three types of brittle features

843 reported here (**Table 1**; see also Kuznetsov et al., 1986 and Angiboust et al., 2021). Yet, transient
844 overpressures likely triggered the formation of some of the (explosive) breccia features (Type III) within
845 already-tectonized Type I and Type II domains (**Figs.11c,13a,b**). These findings are consistent with the
846 episodic opening of vein systems and filling of voids by omphacitic compositions documented by Garcia-
847 Casco et al. (2009) and Cardenas-Parraga et al. (2012) in the Sierra del Convento serpentinite mélange (E.
848 Cuba), and also in other subduction HP-LT mélanges environments (e.g., Muñoz-Montecinos et al.,
849 2021b).

850 *9.5 Insights on slip properties in a fluid-saturated mantle fault zone*

851 Brecciated rocks have long been recognized as potential markers of seismic deformation (e.g.,
852 Sibson, 1986; Angiboust et al., 2012; Melosh et al., 2014). Evidence of fault-zone rocks with diagnostic
853 elements such as damage zones and fault cores (**Fig.13a, b**) highlights for the first time that jadeitites
854 (and associated dark amphibole-rich granofels) host abundant brittle deformation events, most likely
855 along fast-slipping seismogenic fault planes. Estimating slip or strain rates along paleo-fault surfaces (or
856 shear zones) is a challenging task, subject to great uncertainties. While Oncken et al. (in revision) have
857 recently demonstrated that foliated cataclasites can form in slow slip environments for strain rates in the
858 range 10^{-3} to 10^{-5} s⁻¹, the lack of well-preserved pseudotachylytes in the studied samples hampers a
859 direct identification of seismic slip rates (i.e., on the order of m/s). However, the discovery of a foliated
860 fault zone with abundant “shards” exhibiting structures analogous to devitrified spherulites (**Fig.9g**)
861 opens the possibility for constraining former slip events at strain rates approaching seismic slip
862 velocities, fast enough to generate co-seismic temperatures as high as 1500 °C (e.g., Sibson et al., 2006;
863 Menant et al., 2018) and trigger frictional melting. Note that chemical modification of the system is
864 required to explain the enrichment in Ca, Fe and Mg of the shard-bearing fault zone as shown in **Fig.9b**,
865 **c**. As earlier stated by Magloughlin (1992) and Swanson (1992), cataclasites and pseudotachylytes may
866 be closely intricated in fault zone rocks, both exhibiting evidence for open-system modification of the

867 pristine host composition. Thus, the structures herein observed may have formed through (i)
868 metasomatic fault-zone alteration associated with fluid influx within a finely-crushed cataclastic domain
869 (likely a structure as shown in **Fig.13b**), followed by (ii) fast-slip along narrow zones where local melting
870 of the previously-formed cataclasites occurred (thus explaining the enrichment in omphacitic component
871 of “shards” and devitrified spherulites, **Fig.9d-h**). Subsequent shear deformation contributed to the
872 apparent banding systematically observed in Type II fault rocks (e.g., **Fig.9b**). It is therefore proposed
873 that the reported features can be explained by fluctuations in slip rate velocities in a fluid-saturated fault
874 zone.

875 *9.6 Implications for mantle wedge seismicity and plate-interface rheology*

876 Observations made on the Myanmar and Guatemalan loose samples are challenging to link to
877 their respective formation context because pristine field relationships were obliterated during block-in-
878 matrix dismembering as well as sedimentary transport into conglomerate deposits. A few exceptional
879 observations in Myanmar document the presence of jadeitite-bearing “dykes” that are several hundred
880 meters long and meters to tens of meters thick (e.g., Bleek, 1908), but most of these bodies have been
881 mined. The Polar Urals Pus’yerka locality (and to a lesser extent other jadeitite localities in Siberia;
882 Dobretsov & Ponomareva, 1968) has the potential to yield in situ information on the dimensions of
883 ruptured bodies as well as refining the chronology of the processes at stake. Kuznetsov et al. (1986)
884 report that jadeitite “dykes” in the Pus’yerka locality (also mined in the 1990s and mostly exhausted)
885 were locally up to several tens of meters thick, pinching out to meter-thick bands at the northern and
886 southern terminations of the several kilometer-long dyke structure (**Fig.1b**). These observations provide
887 critical insights into the minimum size for mantle wedge earthquakes. Assuming a planar geometry for a
888 single event that ruptured the Pus’yerka jadeitite body (before the dyke structure was boudinaged), a
889 squared 3x3 km rupture, a 10 cm co-seismic displacement and a shear modulus of 84 GPa for jadeitite
890 (Hao et al., 2020), a magnitude of $M_w = 4.5$ can be calculated (Sibson, 1989). This magnitude estimates

891 ranges between 3 and 5 when changing fault dimensions and displacements within bounds compatible
892 with field and structural observations.

893 Is this range of magnitudes comparable with the events reported in the forearc lithospheric
894 mantle clusters in active subduction zones? Davey & Ristau (2011) report a subvertical, 10 km-large
895 cluster of earthquakes (average maximum magnitude of 4.5 for normal, inverse and strike-slip events)
896 above the subduction interface at 40–50 km depth in the northern Hikurangi margin. These authors note
897 their spatial relationship with the inferred 700 °C isotherm, where they expect full antigorite breakdown
898 and relate these earthquakes to dehydration embrittlement processes. Released upwelling fluids
899 infiltrate the mantle wedge, possibly explaining volcanism in the above back-arc basin. Nakajima &
900 Uchida (2018) have identified a nest of supra-slab micro-seismicity ($M_w= 1-3$) located 10-15 km above
901 the Philippine plate in central Japan, between 25–35 km depth with focal mechanisms corresponding to
902 normal and strike-slip events. Interestingly, seismic activity markedly increased after the 2011 Tohoku
903 megathrust rupture, leading Nakajima & Uchida (2018) to consider these events as marking the trace of
904 highly pressurized fluids, cyclically drained above the plate interface along a “highly fractured, pre-
905 existing mature pathway”. According to this model, several months are required for these fluid pulses to
906 travel the few kilometers that separate these nests from the megathrust area (see also White et al.,
907 2016). Similar structural conclusions were drawn by Yu and Zhao (2020) who focused on 35–55 km depth
908 microseismicity events in the NE Japan forearc mantle using P-wave velocity perturbations. Halpaap et
909 al. (2019) have demonstrated that several seismicity nests ranging in magnitude from 2–4 occur under
910 Western Greece in a weakly serpentinized mantle. These events (mostly extensional) occur between 45–
911 55 km depth along a diffuse structure that dips at 45° to the subduction interface plane (see hypocenter
912 distribution plotted in **Fig.14e**). Halpaap et al. (2019) conclude that plate-interface fluids transiently
913 migrate along fault-bounded vents, diminishing the pore pressure along the interface, thus decreasing
914 plate-interface seismicity. This non-exhaustive list of studies confirms that fluids are unequivocally

915 required to explain the seismotectonic features documented in the remote mantle wedge. It also
916 appears that these nests are transient in the context of subduction zone evolution and spatially localized
917 within well-defined structures.

918 The formation of mantle wedge earthquakes has critical implications for plate-interface rheology and
919 more specifically may provide precious constraints on the volumetric abundance of serpentinite in the
920 forearc mantle. A vibrant debate exists on the abundance and distribution of serpentinite in the mantle
921 wedge based on the interpretation of seismic wave velocity perturbations and the effect of serpentinite
922 anisotropy (e.g., Hyndman & Peacock, 2003; Liu et al., 2020; Luo & Wang, 2021). It is likely that
923 serpentinite distribution is highly uneven in the mantle wedge, with the highest serpentinitization rates
924 directly above the interface and fairly low rates (likely less than 15 vol.%) within the innermost forearc
925 mantle (e.g., Reynard, 2013; Abers et al., 2017; Halpaap et al., 2019). Stress accumulation in the mantle
926 must be inhibited by serpentinite formation, which is known to flow at very low stresses (e.g., Escartin et
927 al., 2001; Reynard, 2013 and references therein). The stress level required for enabling plastic
928 deformation of jadeitites (40–125 MPa according to our calculations) is several orders of magnitude
929 greater than the stresses required for the flow of serpentinites (approximately 1 MPa at 600 °C using the
930 calibration from Hilairet et al., 2007). In order to transfer plate-interface stresses into the overlying
931 mantle and trigger plastic deformation of jadeitites (**Fig.4**), the host peridotite must have been relatively
932 dry (i.e., with serpentinitization rates far lower than 10 vol.%) as inferred for many active subduction
933 settings (e.g., Abers et al., 2017; Halpaap et al., 2019). The triggering of multiple cataclastic events as
934 documented by our observations (**Fig.9b**) could be explained by transient increases in strain rate (during
935 slow slip events or earthquakes), by an increase of pore fluid pressures, or by a combination of both. The
936 latter option is supported by the discovery of fabrics typically observed in faulted jadeitites (e.g., **Figs.9,**
937 **12**), as well as major and trace element chemical arguments that point to mass transfer and opening of
938 the system to external fluids (e.g., **Fig.6**), most likely sourced in the subduction interface region.

939 From these considerations, we can conclude that (i) the seismic events recorded above the subduction
940 interface in several active subduction zones likely correspond to the brittle deformation of jadeitite
941 bodies (**Fig.14d,e**), which according to field observations are themselves derived from the metasomatic
942 replacement of former felsic dyke bodies or direct precipitation from fluids (see discussion in Harlow et
943 al., 2015 and Angiboust et al., 2021); and (ii) brittle jadeitite deformation may have occurred in the fossil
944 subduction record and in present-day settings in a relatively dry peridotitic mantle, thus providing
945 valuable insights into the serpentinization degree of the interface hangingwall and the stress distribution
946 at the base of the megathrust region (Halpaap et al., 2019; Luo & Wang, 2021). This record of the deep
947 conduits draining plate-interface fluids into the overlying mantle wedge sheds light on the structure of
948 the “cold nose” above the subduction interface, with implications for volatile budget estimates, the
949 rheology of the plate interface itself (including the various types of seismicity), and the interpretation of
950 V_p/V_s and Poisson’s ratios from active subduction settings worldwide.

951 **10. Conclusions**

952 A vein network from the Polar Urals composed of jadeitite and amphibole-phlogopite granofels is hosted
953 in mildly serpentinized peridotites. This structure formed by metasomatic replacement of a former
954 magmatic dyke above the subduction interface at c. 50 km depth, in a young and dry mantle wedge
955 environment. During subsequent cooling of the subduction zone, this jadeitized “dyke” experienced
956 mixed brittle-viscous deformation regimes associated with several serpentinite- and oceanic crust-
957 derived fluids. Extensive fracturing, well-visible in the amphibole-rich granofels and more cryptic in the
958 jadeitite bodies, led to the local formation of breccias and foliated cataclasites. The associated
959 metasomatic imprint confirms the importance of these fault zones as major drains for overpressurized
960 plate-interface metamorphic fluids. Very similar features, including omphacite- and amphibole-rich
961 breccias together with foliated cataclasites, are also observed in loose jadeitite boulders from Myanmar
962 and Guatemala serpentinite mélanges. We propose that supra-slab seismicity nests recorded in present-
963 day subduction margins (e.g., N Hikurangi, NE Japan, W Greece) may reflect the repeated fluid-assisted
964 breakage of planar bodies formed by jadeitite “dykes”. The local presence of these rupture events
965 implies that the serpentinization rate was locally low (probably <10–15 vol.%) along the base of the

966 mantle wedge during “dyke” faulting activity. These results shed light on the importance of physical
967 discontinuities such as jadeitite bodies for draining plate-interface fluids into the mantle wedge, and
968 provide key constraints on the stress distribution in deep subduction margins.

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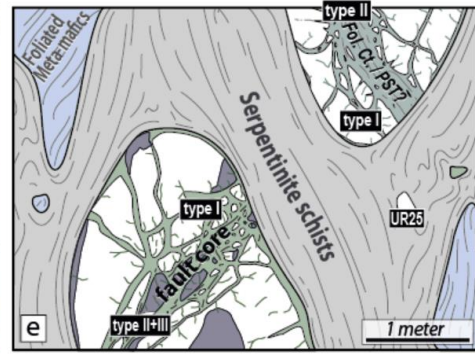
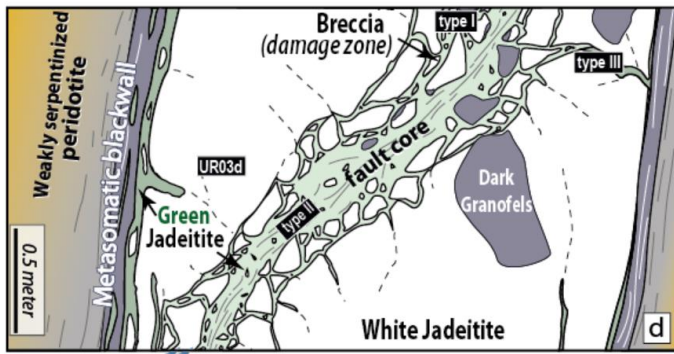
1379

1380 **Table 1.** Summary of selected samples for this comparative study, including their paragenesis and the
1381 various fracturing patterns identified therein (see text for details on the criteria used for this
1382 classification). Numbers in the last three columns refer to the chronological sequence of fracturing events
1383 identified in each sample.

1384

	Sample	Region	Pre-fracturing assemblage	Paragenesis associated w/ brittle def.	Fracturing pattern		
					type I	type II	type III
Polar Urals	PU2	Pus'yerka	white Jd (Jd ₁), Ed	Omp, Mg-ktp, Rct, Eck, Phl, Clc	x (1)		x (2)
	PU5	Pus'yerka	white Jd (Jd ₁)	Omp, Phl, Clc (±Cal)		x	
	UR11b	Pus'yerka	Ed or Mg-ktp with Omp inc.	Jd ₂ , Omp, Mg-ktp, Eck/Nyb, Ttn, Chr	x		
	UR03b,03d	Pus'yerka	white Jd (Jd ₁)	Jd ₂ , Omp			x
	UR25	Nephrite brook	white Jd (Jd ₁)	Jd ₂ , Omp, Ttn	x		
Guatemala	MTG00	Motagua region	white Jd	Omp (<i>not analyzed</i>)	x		
	MTG01	South Motagua	white Jd (Jd ₁)	Jd ₂ , Omp, Lws, Ttn	x		
	MTG02	South Motagua	white Jd (Jd ₁)	Jd ₂ , Omp, Ttn			x
	MTG03	South Motagua	white Jd (Jd ₁)	Jd ₂ , Omp, Ttn (±REE-rich Ep)		x	
	MTG12	South Motagua	white Jd (Jd ₁)	Jd ₂ , Omp, Ttn		x	
Myanmar	KAS01	Hpakan (Kashin)	white Jd (Jd ₁)	Jd ₂ , Omp, Eck	x (1)		x (2)
	KAS04	Hpakan (Kashin)	white Jd (Jd ₁)	Jd ₂ , Omp, Eck, Clc (±Cls)	x (1)	x (2)	x (3)
	KAS06	Hpakan (Kashin)	Ed (+Omp?)	Jd ₂ , Omp, Rct, Eck	x		
	KAS07	Hpakan (Kashin)	white Jd (Jd ₁) and Mg-ktp	Jd ₂ , Omp, Eck, Chr		x (1)	x (2)
	KAS10	Hpakan (Kashin)	white Jd (Jd ₁)	Jd ₂ , Eck		?	

Table 1



H₂O, Ca, Fe, Mg
 (+LILEs, ±Cr, ±Ti ±REEs)

Fault-forming environment

Present-day occurrence in serpentinite mélanges