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Title: Seismic properties of mantle metasomatism from mantle xenoliths beneath the North Tanzania Divergence, East African Rift

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Key words: mantle xenolith, seismic properties, alkaline magma, mantle metasomatism, rift initiation, East African Rift

Highlights:

- Mantle at rift initiation is metasomatized by alkaline fluids at 40-90 km depth
- 1100°C mantle isotherm relates to the volatile-bearing peridotite solidus
- Lithospheric mantle has ≈20% of crystallized or ≈10% of fluid-filled vertical veins

1 Abstract

2 We use mantle xenoliths brought to the surface by alkaline lavas to determine the
3 chemical and physical properties of the metasomatized lithospheric mantle that contribute to
4 the earliest rifting stage in East Africa. Our results help to interpret the seismic tomographic
5 images in terms of vein and inclusions proportions in the lithospheric mantle. We focus on
6 mantle xenoliths from the in-rift Pello Hill volcano in the North Tanzanian Divergence (NTD).
7 These xenoliths reveal the presence of refractory mantle harzburgites and dunites with coarse
8 granular to porphyroclastic textures and 6 to 80% of diopside, phlogopite and amphibole-
9 bearing veins and phlogopite-rich hornblendite xenolith. The presence of calc-potassic and
10 FeO, TiO₂-rich veins, and mineral equilibria of olivine and pyroxenes indicate that fluid/melt-
11 rock interactions occurred at depth from 40 km to 80-90 km, and indicate the presence of a
12 high-temperature isotherm beneath the NTD (T=1040-1200°C). We computed the seismic
13 properties of the mantle xenoliths with different proportions, compositions, and geometric
14 distributions of crystallized and fluid-filled veins. Compared to vein-free peridotites, for
15 crystallized vein-bearing xenoliths, the velocity is lowered by 2-4% to 28-37% for V_p and by 2-
16 3% to 25-29% for V_s for 6% to 60% veins, respectively. For fluid-filled inclusions, hydrous melt
17 lens-shape inclusions are the most effective parameter to reduce P velocity, compared to dry
18 or 2.5%-CO₂ peridotitic melt. A comparison with seismic tomography velocities allows us to
19 discuss the current state of the lithospheric mantle. The best agreement obtained between P
20 teleseismic tomography (V_p anomalies between -9% and -15%) and vein-bearing peridotites
21 (depth 40-90 km) corresponds to 12-25% of crystallized veins or 8-15% for fluid filled-veins for
22 a vertical foliation and transtensional strain regime in the mantle lithosphere beneath the
23 NTD.

24

25 1 Introduction

26 A remaining challenge for the geophysics and rock physics communities is to separate
27 the different factors affecting seismic properties such as pressure, temperature,
28 compositional variations, fluid presence or mantle anisotropy (Almqvist and Mainprice, 2017;
29 Deschamps and Trampert, 2003; Mavko et al., 2020). Partial melt signature is particularly
30 problematic as different factors may produce similar decreasing effects on seismic velocities
31 (temperature increase, change of shape or increase of size of melt pockets, fluid presence,
32 anisotropy...) (Almqvist and Mainprice, 2017; Deschamps and Trampert, 2003; Mainprice,
33 1997). Laboratory experiments and models have been extensively carried out to quantify
34 these melt effects on seismic velocities and have pointed out a non-linear relationship with a
35 strong dependency on the shape and orientation of the melt pocket (Clark and Leshner, 2017;
36 Goes et al., 2000; Hammond and Kendall, 2016). Seismic velocity also relates to intrinsic
37 properties of rocks (from minerals) and extrinsic components (open cracks, crystallized and
38 fluid-filled veins, alteration) (Almqvist and Mainprice, 2017). Mantle metasomatism is one
39 cause for a change in the seismic properties of the rocks, as it affects both their intrinsic and
40 extrinsic components (Dawson, 1984; O'Reilly and Griffin, 2013). The percolation of
41 anhydrous, hydrous, or carbonatitic silicate melts overprints the mantle lithosphere signature.
42 It modifies the viscoelastic and chemical properties by forming compositionally distinct
43 secondary phases (interstitial minerals, metasomatic veins) or by modifying minerals
44 equilibrium (modal and cryptic metasomatism, respectively, Dawson, 1984; O'Reilly and

45 Griffin, 2013). The presence of new metasomatic minerals, such as amphibole or phlogopite,
46 may in turn, affect the seismic properties of the metasomatized mantle. When metasomatism
47 is related to a mantle plume, it also alters the geotherm and, consequently, the seismic
48 velocities observed in tomographic models (e.g., Allen et al., 2002; Grijalva et al., 2018).

49 In this study, we aim to quantify the effect of metasomatism on lithospheric seismic
50 velocities. We take advantage of xenoliths collected in the East African rift, in the so-called
51 North Tanzanian Divergence (NTD). Here, the Archaean sub-continental lithospheric mantle
52 has experienced multiple metasomatic events that have modified its composition and
53 rheology (Dawson, 1997; Koornneef et al., 2009; Lee and Rudnick, 1999). In particular, the
54 most recent impingement of the East Africa mantle plume has modified the geotherm and
55 also carried an interaction of volatile-rich melts or fluids with peridotitic mantle (Baptiste et
56 al., 2015; Gibson et al., 2013; Mulibo and Nyblade, 2013a; Reiss et al., 2021; Selway et al.,
57 2014). In order to evaluate the role of the metasomatic agent on the lithospheric mantle, we
58 perform petrological and petrophysical investigations for vein-bearing, and vein-free mantle
59 xenoliths from Pello Hill in the NTD to: (1) characterize the chemistry, and the depth of the
60 melt/fluid percolation within the lithospheric mantle; (2) determine the seismic properties of
61 the metasomatized mantle; and (3) estimate the proportion of fluid circulating within the
62 lithospheric mantle by comparing seismic properties of mantle xenoliths and P-wave
63 tomographic images. In this study, we evaluate the link between geophysical and petrological
64 data based on: (1) observations made on natural samples considered representative for the
65 lithospheric mantle beneath the NTD; and (2) constraints from modeling based on
66 experimental mineral and rock physics laboratory measurements.

67

68 **2 The North Tanzanian Divergence**

69 The North Tanzanian Divergence is located at the southern tip of the Eastern branch
70 of the East African Rift. This zone is a relatively young continental rifting zone (~8 Ma, Le Gall
71 et al., 2008), compared to the Afar mature zone (45-35 Ma, e.g., Courtillot et al., 1999;
72 Courtillot and Renne, 2003; Ebinger and Sleep, 1998; George et al., 1998) and represents the
73 initial stages of the rifting process. In the NTD, the complex interaction between inherited
74 structures in the highly heterogeneous lithosphere and deep mantle processes leads to diffuse
75 tectonic deformation and magmatism at the surface (Huerta et al., 2009; MacDonald et al.,
76 2001; Mulibo and Nyblade, 2013b; O'Donnell et al., 2013; Pik et al., 2006; Tiberi et al., 2019;
77 Weeraratne et al., 2003). The narrow, poorly magmatic Kenya rift abruptly widens when
78 reaching North Tanzania (~3° S) and splits into three different arms: the Eyasi, Manyara and
79 Pangani branches (Figure 1, Corti et al., 2013; Le Gall et al., 2008). These branches develop
80 mainly in Proterozoic mobile belts, at the edge of (or in) cratonic units (Tanzanian Craton and
81 Masai block) with magmatism occurring since ~8-5 Ma (Dawson, 1992; Mana et al., 2015), and
82 a tectonic expression since ~3 Ma (Foster et al., 1997). The volcanism preceded the faulting
83 by ~20 Ma in the Afar (Ebinger et al., 2000), and by a few Ma in the Kenyan rift and in the NTD
84 (Le Gall et al., 2008; Mechie et al., 1997).

85 The North Tanzania Divergence volcanism is related to the presence of a mantle
86 plume (Adams et al., 2012; MacDonald et al., 2001; Nyblade et al., 2000). The NTD displays
87 broadly distributed volcanic edifices with a highly heterogeneous chemical signature (Dawson,

88 2008; Mana et al., 2015). It encompasses the oldest pre-rift volcano Essimingor and the only
89 active carbonatitic edifice, Oldoinyo Lengai. On a N-S axis, in-rift volcanism is highly alkaline
90 (e.g., Baudouin et al., 2016; Baudouin and Parat, 2020; Klaudius and Keller, 2006; Lee and
91 Rudnick, 1999) and sub-alkaline to alkaline for the E-W axis (Braunger et al., 2021; Mana et
92 al., 2015, 2012; Mollel et al., 2008; Nonnotte, 2007; Roberts, 2003; Wilkinson et al., 1986)
93 (Figure 1). This diversity of magmas results from variable amount (3-6% to 1%, Baudouin and
94 Parat, 2020; Mana et al., 2015) and depths of partial melting. From north to south of the NTD
95 the depth of partial melting increases, from 75-90 km in the Natron-Engaruka basins
96 (Mattsson et al., 2013), and up to 110-130 km for Kwahara and 150 km for Labait volcanoes
97 (Baudouin and Parat, 2020; Gibson et al., 2013).

98 Although the magmatism strongly highlights the presence of recent mantle
99 metasomatism by alkaline melts, previous Re/Os and Nd/Sm isotope studies indicate the
100 presence of ancient events, during the Precambrian or Pan African orogens (Burton et al.,
101 2000; Chesley et al., 1999; Koornneef et al., 2009). The records of recent and ancient
102 metasomatic equilibrium indicate that the lithospheric mantle below the NTD is highly
103 heterogeneous and that long-term metasomatism is still present. In the NTD, five volcanoes
104 erupted mantle xenoliths: Eledoi, Pello Hill, Lashaine, Olmani, and Labait (Figure 1). Combined
105 with information from the magmatism, these xenoliths give an insight of the past and almost-
106 current mantle state at depth, and allow an estimate of the impact of metasomatism on
107 seismic properties.

108

109 **2.1 Lithospheric mantle metasomatism beneath the NTD**

110 At the NTD, the strong and thick (150-180 km) cratonic lithosphere of the Tanzanian
111 craton highly contrasts with the thinner and younger Proterozoic mobile belt lithosphere
112 (Figure 1, Craig et al., 2011; Fishwick, 2010; Lee and Rudnick, 1999; Tiberi et al., 2019). The
113 roughly N-S suture between these units is thought to have played a first-order structural
114 control on the evolution of the rift (Corti et al., 2013; Ebinger et al., 1997; Koptev et al., 2018,
115 2015). Pre-existing zones of weakness could have prevented the cratonic lithosphere from
116 being deformed by accommodating external stresses (Lenardic et al., 2003). It may also have
117 channelized the mantle upwelling and generation of melts and/or deep fluid-rock interactions
118 inducing modal and/or cryptic metasomatism (Albaric et al., 2009; Baker et al., 1998; Delpech
119 et al., 2004; Dixon et al., 2008; Grégoire et al., 2000; Koptev et al., 2016; O'Reilly and Griffin,
120 1988; Wölbern et al., 2012).

121

122 **2.1.1 Cratonic mantle**

123 The Tanzania cratonic lithospheric mantle, down to ~120 km, has a mean Re-Os age
124 of 2.8-2.0 Ga (xenoliths from the on-craton border Labait volcano, Chesley et al., 1999),
125 suggesting that refractory lithospheric mantle has been metasomatized during a 2.8 Ga melt
126 event. Nd ages of 2.1-1.76 Ga, associated with a Light Rare Earth Elements enrichment in
127 clinopyroxenes from spinel or garnet-bearing peridotites, may indicate cryptic metasomatism
128 induced by subduction-related fluid/melt during the Usagaran belt formation (Koornneef et

129 al., 2009). Deeper garnet-bearing xenoliths (> 120 km) have Re-Os model ages < 1 Ga and are
130 younger than the spinel-bearing xenoliths, suggesting a more recent melting event that
131 overprinted the 2.8 Ga event (Chesley et al., 1999). This is also confirmed by Nd model ages
132 of 698-639 Ma, associated with hydrous fluids, possibly coming from subduction related to
133 the Pan-African Orogeny and/or to the plume ascension and rift opening (Hui et al., 2015;
134 Koornneef et al., 2009). The north Tanzania lithospheric mantle is therefore heterogeneous,
135 with at least two metasomatic events, resulting from asthenospheric Fe-rich melts (Lee and
136 Rudnick, 1999) and silicic fluid circulation (Aulbach et al., 2011). The mantle below Labait has
137 an isotropic composition intermediate between the Depleted Mantle (DM) and Enriched
138 Mantle 1 (EM1) reservoirs that has been associated with former subduction (Aulbach et al.,
139 2011). The occurrence of deep metasomatic processes is also supported by the microstructure
140 of xenoliths, where neoblasts crystallized (due to temperature increase) in the deepest
141 garnet-free and uppermost garnet-bearing xenoliths (Vauchez et al., 2005).

142

143 **2.1.2 In-rift mantle**

144 Below the Masai block and the East-West volcanic axis (in-rift xenoliths from Lashaine
145 and Olmani volcanoes), the lithospheric mantle is at least 3.4 Ga old (Burton et al., 2000) and
146 was also subject to different metasomatic events. A melting event was dated by U-Pb for
147 Lashaine garnet-bearing xenoliths at 2.0 Ga, and Re-Os analyses on silicate minerals from
148 garnet-bearing Lashaine xenoliths display ages of 15.4-31.4 Ma (Burton et al., 2000; Cohen et
149 al., 1984). The garnet-bearing peridotites have higher Al₂O₃, CaO and Na₂O concentrations
150 than garnet-free peridotites, but both have similar Mg# (Gibson et al., 2013; Rudnick et al.,
151 1994). These compositions were interpreted reflecting fluid/melt derived from subducted
152 oceanic crust, and the EM1 and EM2 reservoirs as detrital sediment component in the hotter
153 Archean mantle (Aulbach et al., 2011; Rudnick et al., 1994). The metasomatism in Olmani
154 xenoliths is marked by the presence of clinopyroxene and phosphates, suggesting an
155 interaction of the peridotites with carbonatitic melt (Rudnick et al., 1994).

156 Below the Pello Hill volcano, the lithosphere was dated to at least 1.3 Ga by Re-Os isotopes
157 for a lherzolite, and Os model ages might indicate an Archean lithosphere (Aulbach et al.,
158 2011). A negative correlation of Re-Os isotopes evolution may suggest that the lithospheric
159 mantle has been metasomatized less than 16 Ma ago by melts with silicic and carbonatitic
160 intermediate compositions (Aulbach et al., 2011; Burton et al., 2000). In addition, isotopic
161 signatures for Ol Doinyo Lengai and the Ngorongoro volcanic complex are evidence of even
162 more diverse enriched mantle signatures (HIMU and EM1, Aulbach et al., 2011; Bell and
163 Simonetti, 1996; Mollet et al., 2011, 2009).

164

165 **3 Methods**

166 **3.1 Samples**

167 To overview the lithospheric mantle metasomatism below the in-rift Pello Hill
168 volcano, we selected a representative set of 12 xenoliths with and without veins, and with

169 variable amounts of interstitial metasomatic minerals Table 1. Among those 12 samples, 7
170 samples are free of veins (PH1, PH2, PH8, PH13, PH17, PH24, PH26) and 4 samples have 6 to
171 80% of clinopyroxene-phlogopite-amphibole veins crosscutting the xenoliths (PH4, PH9, PH18
172 and PH27). The host xenoliths are dunites (PH4, PH8, PH9, PH18 and PH26) and harzburgites
173 (PH1, PH2, PH13, PH17, PH24 and PH27), plus 1 phlogopite-rich, olivine-, and clinopyroxene-
174 bearing hornblendite (PH23). We made 12 thin sections, perpendicular to the veins if present,
175 in order to have a representative vein proportion of the mantle xenoliths and to perform
176 EPMA and EBSD analyses.

177

178 3.2 Pressure and temperature estimate

179 Major element concentrations of minerals were measured at the Paul Sabatier
180 University of Toulouse, with a Cameca SXFive electron microprobe (EPMA). The analyses were
181 carried out with an accelerating voltage of 15 kV, a 10 or 20 nA beam current depending on
182 the mineral resistance under the beam (10 nA for amphibole and phlogopite; 20 nA for olivine,
183 pyroxene, garnet, and oxides), a focused beam (inferior to the μm) for olivine, clinopyroxene,
184 orthopyroxene, garnet, oxide and defocused (2 μm) for amphibole and phlogopite. The
185 analyzed surface is about $2 \times 2 \mu\text{m}^2$. The counting time was fixed to 10 seconds for all elements
186 except for fluorine which was set to 20 seconds. The standards used for calibration were albite
187 for Na, wollastonite for Si and Ca, Al_2O_3 for Al, sanidine for K, MnTiO_3 for Mn and Ti, Fe_2O_3 for
188 Fe, topaz for F, MgO for Mg, tugtupite for Cl, BaSO_4 for Ba, Cr_2O_3 for Cr, SrSO_4 -MAC for S,
189 native metal for Cu and Ni (Ni-G5), and sphalerite for Zn.

190 We estimate the equilibration depth and temperature of mantle xenoliths using
191 orthopyroxene-clinopyroxene and olivine-clinopyroxene equilibria that have been calibrated
192 experimentally by Brey and Köhler (1990) and Köhler and Brey (1990), respectively. The
193 equilibria between orthopyroxene-clinopyroxene and olivine-clinopyroxene were tested by
194 using Fe/Mg partitioning between minerals ($\text{Fe}^{2+}/\text{Mg} \times 100$, Brey and Köhler, 1990; Putirka,
195 2008, and Supplementary Material 1). For all mineral equilibrium, we chose grains as close as
196 possible to each other, and far from veins. By using the two-pyroxenes thermometer from
197 Putirka (2008), we determined an a priori equilibrium temperature (T_{input}). Then, an a priori
198 pressure (P_{input}) is determined graphically from the geotherm modified by a plume thermal
199 anomaly from Selway et al. (2014) and T_{input} . Finally, T_{input} and P_{input} were implemented in
200 PTEXL3 spreadsheet (from T. Koehler and modified by A. Girnis; [http://www.mineralogie.uni-](http://www.mineralogie.uni-frankfurt.de/index.html)
201 [frankfurt.de/index.html](http://www.mineralogie.uni-frankfurt.de/index.html)), where the final temperatures is determined with the
202 orthopyroxene-clinopyroxene equilibrium from Brey and Köhler (1990). In the case of
203 xenoliths without orthopyroxene, the partitioning of Ca between the olivine and the
204 clinopyroxene from Köhler and Brey (1990) was used. The final pressure was computed using
205 Ca partitioning between the olivine and the clinopyroxene from Köhler and Brey (1990) for
206 the garnet-free mantle xenoliths from Pello Hill.

207 The calcium concentration in olivine was measured with EPMA, and several analyses
208 were confirmed by Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS)
209 at AETE-ISO platform (OSU OREME, University of Montpellier). A 130 μm diameter laser beam
210 was used for olivine with a pulsed 193 nm Excimer CompEx 102 laser (GeoLas Q+ platform).
211 The laser repetition rate was 6 Hz and the laser power was $5.75\text{-}6 \text{ J}\cdot\text{cm}^{-2}$. Concentrations were

212 calibrated with the intern standard NIST612, and SiO₂ concentrations were previously
 213 measured by the electron microprobe. The external standard is BIR. The raw laser data were
 214 processed with GLITTER software (Griffin et al., 2009), which converts the signal (signal
 215 intensity/time) to elemental concentrations. For thermobarometry, the pressure-
 216 temperature (P-T) conditions were computed using both microprobe and LA-ICPMS analyses.

217

218 **3.3 Crystallographic preferred orientation**

219 We analyzed ten samples with the Camscan Crystal Probe X500FE Electron Back
 220 Scattered Diffraction (EBSD) microscope at Geosciences Montpellier (MEA platform,
 221 University of Montpellier) to determine the crystallographic preferred orientation (CPO) and
 222 phase proportion of olivine, clinopyroxene, orthopyroxene, phlogopite, amphibole, and
 223 oxides. Measurements were performed using an acceleration voltage of 20 kV and an average
 224 working distance of 25 mm. For each sample, the map covers almost the entire thin section
 225 (30×40 mm). The step size between hitting points varies from 12 to 40 μm depending on the
 226 sample. The indexing rate varies between 84% (PH23, olivine-clinopyroxene-phlogopite-
 227 amphibole-oxide vein) and 96% (PH26 dunite), depending on the mineral species and the
 228 degree of fractures in the sample. The lowest indexing rates are for samples containing a high
 229 proportion of phlogopite, which is difficult to index.

230 **3.4 Crystallographic textures and seismic properties**

231 **3.4.1 Crystallographic textures**

232 The MTEX toolbox has been used to determine and represent the crystallographic
 233 textures and the J-index from EBSD data (Hielscher and Schaeben, 2008; Mainprice et al.,
 234 2011). The J-index quantifies the fabric strength and is dimensionless. It results from the
 235 integration of the density of the Orientation Distribution Function (ODF, $f(g)$) at a g orientation
 236 over the entire volume (Ismail and Mainprice, 1998):

$$237 \quad J = \int f(g)^2 dg \quad (1)$$

238 where $dg = d\varphi_1 d\phi d\varphi_2 \sin\phi / 8\pi^2$ and φ_1 , ϕ and φ_2 are the Euler angles, which represent the
 239 orientation of a crystal within the sample reference frame. $J = 1$ for a random distribution of
 240 the crystallographic axes, while J is infinite for single crystal orientation.

241

242 **3.4.2 Seismic properties**

243 We use the AnisEulerSC program to model the seismic properties of the Pello Hill
 244 mantle xenoliths. It allows us to consider the presence of different mineral phases, the
 245 crystallographic preferred orientations, and the presence of fluid and melt inclusions (Kim et
 246 al., 2020). AnisEulerSC is an extended version of the MTEX program (Hielscher and Schaeben,
 247 2008; Mainprice et al., 2011) which uses the CPO of the major phases (measured with EBSD),
 248 their respective modal content, and the stiffness tensor for each phase. In addition, the
 249 AnisEulerSC program considers the grain mean aspect ratio and the grain volume for the

250 seismic property computation (Kim et al., 2020) (see Supplementary Material 2 for
 251 comparison between MTEX and AnisEulerSC velocities for two very contrasting compositions).
 252 It is important to note that the elastic tensor of the aggregate varies as of function of the grain
 253 aspect ratio (Koch et al., 2004). AnisEulerSC also offers the possibility to consider oriented
 254 cracks that can be filled with crystallized minerals or fluids. It is based on the introduction of
 255 elastic inclusions into a background medium and the calculation of the average Self-Consistent
 256 (SC) elastic tensor (Eshelby, 1957; Hill, 1965) from the individual stiffness tensor of the fluid
 257 or the different minerals in the aggregate. For this method, each component is considered as
 258 an isolated inclusion in an infinite isotropic homogeneous background medium (Berryman and
 259 Berge, 1993).

260 For our study, we consider the stiffness tensors at a confining pressure of 2 GPa and
 261 ambient temperature. The 2 GPa pressure matches the mean depth (between 40 and 90 km
 262 depth) of the mantle xenoliths used in this study. The temperature effect was not
 263 implemented in the stiffness tensor due to the lack of experimental temperature derivatives
 264 for phlogopite and amphibole. The stiffness tensors used in the calculations are from
 265 Abramson et al. (1997) for olivine, Sang and Bass (2014) for diopside (clinopyroxene), Chai et
 266 al. (1997) for orthopyroxene, Peng and Mookherjee (2020) for amphibole, Chheda et al. (2014)
 267 for phlogopite and Da Silva et al. (1999) for ilmenite. The magnetite stiffness tensor of
 268 Reichmann and Jacobsen (2004) was used instead of chromite (which was indexed as
 269 magnetite in EBSD) because we found no pressure derivatives for chromite in the literature.
 270 Moreover, the chromite phase is present in a small amount, so the expected impact on the
 271 calculated velocities should be negligible. We corrected the effect of temperature on
 272 velocities for vein-free and crystallized vein-bearing peridotites aggregate (Table 1 and Table
 273 2, respectively) by applying a percentage decrease. This latter was calculated by using the
 274 velocity variation between a peridotite at 2 GPa/ambient temperature and 2 GPa/1200°C. At
 275 1200°C, the mean average velocities decrease is 7% for $V_{p_{max}}$, 9% for $V_{p_{min}}$, 10% for $V_{s1_{max}}$,
 276 and 11% for $V_{s1_{min}}$, $V_{s2_{max}}$ and $V_{s2_{min}}$ (Supplementary Material 3).

277 We first determined the velocities of all the xenoliths (peridotite or peridotite + vein)
 278 by computing the SC tensor in only one step (Figure 4). For the computation, we use the elastic
 279 tensor, a mean measured aspect ratio, and an ellipse approximation for each mineral. To
 280 consider the orientation of veins in the mantle xenoliths, we then determined the velocity by
 281 computing the average SC tensor in two or three steps, considering fluid-filled or crystalline
 282 veins, respectively, for 4 vein-bearing samples (PH4, PH9, PH18 dunites, and PH27
 283 harzburgite). For peridotite with crystalline veins, we compute 1) the vein-free peridotite
 284 average SC tensor, 2) the crystallized vein average SC tensor, and 3) the average peridotite +
 285 vein SC tensor for a given vein orientation. In the final step (3), to obtain only the mean SC
 286 tensor of the peridotite (without orientation), the peridotite minerals are defined as spheres
 287 while we approximate the vein minerals by ellipses. For the fluid-filled veins, we omit the
 288 second step (SC tensor of the vein) as the fluid bulk moduli are known from experimental
 289 density measurements (Genge et al., 1995; Sakamaki et al., 2011, 2010, 2009).

290 Finally, the velocities are calculated in all directions from the average SC tensor using the
 291 equation from (Christoffel, 1877) which relates the elastic stiffness tensor with V_p , V_{s1} , V_{s2} ,
 292 and the anisotropy. The anisotropy (AV) is computed as followed:

$$293 \quad AV = 200 \times \frac{V_{max} - V_{min}}{V_{max} + V_{min}} \quad (2)$$

294 where V_{max} is the maximum velocity and V_{min} the minimum velocity for V_p or V_{s1} or V_{s2} in the
295 sample.

296

297 **4 Pello Hill mantle xenoliths**

298 **4.1 Texture and mineralogy**

299 Pello Hill mantle xenoliths are refractory vein-free and vein-bearing harzburgite and
300 dunite with coarse granular to porphyroclastic textures with subequant to tabular olivine and
301 pyroxene grains (0.1–5 mm) (Table 1, Table 2). Some of the largest olivine crystals are
302 elongated, with shape ratios up to 1:10. Their shape-preferred orientation marks the foliation.
303 These large crystals usually display undulose extinction and subgrain boundaries suggesting
304 deep and intracrystalline deformation (Passchier and Trouw, 2005). Orthopyroxenes (2 to 20
305 modal percent in dunite and harzburgite, respectively) present either exsolution of
306 clinopyroxene, amphibole inclusions, or clinopyroxene growth at rims (Figure 2.a).
307 Clinopyroxenes in peridotites (out of the vein) are scarce, occur as isolate interstitial crystals
308 with curved margins (Figure 2.b), and can be associated with phlogopite and amphibole small
309 interstitial grains, or as inclusions. Phlogopites and/or amphiboles are present as interstitial
310 “patches”, sometimes associated with clinopyroxenes (Figure 2.c and d). Some patches
311 contain spinels with no coronitic reaction. Oxides can be found as inclusions in clinopyroxene
312 and phlogopite (Figure 2.c and d) or interstitial in the peridotite (Figure 2.b). Some interstitial
313 sulfurs are also present within the peridotites.

314 In the vein-bearing xenoliths, the vein proportion ranges from 6% (PH18) up to 100%
315 (PH23) of the thin section (Table 1). The rims of the vein are well-defined with large crystals
316 (Figure 2.f) or diffuse with small interstitial grains (crystals with 500-2100 μm in size in the
317 center of the vein and 80-150 μm in size at the rim; Figure 2.e). Veins comprise an assemblage
318 of clinopyroxene, phlogopite, amphibole, and oxide (Figure 2.e, f and h). The proportions of
319 clinopyroxene, phlogopite, amphibole, and oxide vary from (Table 1): high proportion of
320 clinopyroxene (40-60%) and low proportion of amphibole (12-17%) (PH4 and PH18) to high
321 amphibole (41-52%) and low clinopyroxene (19%) and phlogopite (18-22%) proportions (PH9
322 and PH27). PH23 xenolith is a hornblendite with amphibole (46%), phlogopite (20%),
323 clinopyroxene (17%), and interstitial olivine (16%) (Figure 2.g).

324

325 **4.2 Mineral composition**

326 **4.2.1 Olivine**

327 Olivines are Mg-rich with Mg# ($=\text{Mg}/(\text{Mg}+\text{Fe}^{2+})\times 100$) ranging from 87.7 (PH4) to 93.1
328 (PH18) in vein-bearing peridotites, and from 92.0 (PH26) to 93.1 (PH13) in vein-free
329 peridotites. Olivines from PH18 dunite have a Mg# gradient with Mg# decreasing approaching
330 the metasomatic vein (89.5 near the vein and 93.1 far from the vein). Olivines in veins (PH23)
331 have low NiO and SiO₂ contents (0.1-0.2 wt.% and 38.5-39.7 wt.%, respectively) and low Mg#

332 (77.5 to 79.6) compared to olivine in peridotites (0.3-0.5 wt.% NiO, 40.9-42.1 wt.% SiO₂ and
333 Mg# = 87.7-93.1; Supplementary Material 4).

334

335 **4.2.2 Pyroxenes**

336 Orthopyroxenes, present in most of peridotites, are enstatites (Supplementary
337 Material 4). Orthopyroxenes in the vein-bearing samples have a lower Mg# (<92), i.e. higher
338 concentrations in FeO (> 5 wt.%) and lower concentrations in MgO (< 35 wt.%) compared to
339 vein-free peridotites (Mg# = 92.4-93.8).

340 Clinopyroxenes have diopside/augite compositions (Wo₄₅En₅₁Fs₄ to Wo₃₃En₆₃Fs₄).
341 Clinopyroxenes are diopsides in the veins and vein-bearing peridotites (PH4, PH9, PH18,
342 PH23), and are augite in the vein-free samples (PH1, PH13, PH17, PH26). Compared to
343 clinopyroxene in peridotites, the clinopyroxenes in veins are enriched in CaO (21-23 wt.%,
344 Figure 3.a) and depleted in Cr₂O₃ (< 1 wt.% in vein and 1.5-4 wt.% out of vein, Supplementary
345 Material 4). In the vein-bearing samples (PH4, PH9, and PH18), there is a chemical gradient
346 from the vein to the peridotite from 0.02 up to 3.0 wt.% for Cr₂O₃ and from 22.7 to 19.2 wt.%
347 for CaO (Figure 3.a). In the dunite sample (PH4), the core of clinopyroxenes in the veins have
348 0.03-0.17 wt.% of Cr₂O₃ and 22.0-22.7 wt.% of CaO, whereas near the vein, Cr₂O₃ and CaO
349 concentrations in clinopyroxene core are 2.4-2.5 wt.% and 21.4 wt.%, respectively. In the
350 dunite, far from the vein, clinopyroxenes have 2.9-2.9 wt.% of Cr₂O₃ and 21.2-21.3 wt.% of
351 CaO (Figure 3.a). All clinopyroxenes are unzoned except in dunite sample PH4 with Cr₂O₃
352 ranging from 2.4-3 wt.% in the core down to 1.6-2.1 wt.% at the rims. The vein-bearing dunite
353 PH9 has clinopyroxenes in the peridotite with a large variation in Cr₂O₃ concentrations (from
354 0.7 to 2.4 wt.%). These clinopyroxenes with the lowest Cr₂O₃ values have similar
355 concentrations to clinopyroxenes occurring within the veins (Supplementary Material 4).

356

357 **4.2.3 Amphibole**

358 The amphiboles are calcium-rich (8.8-11.0 wt.% CaO) with Mg# ranging from 77.8
359 (vein sample, PH23) up to 93.4 (vein-free sample, PH13) (Figure 3.b and c). Most of them are
360 edenite (Leake et al., 1997). Only rare amphiboles in veins from sample PH4 and PH23 are
361 classified as pargasite.

362 Amphiboles present in veins or as interstitial minerals in peridotites (further called
363 "patch amphibole") are chemically distinct. Three chemical groups can be defined: i) vein-free
364 samples and a PH18 patch (lowest vein proportion, 6%) contain amphiboles with FeO < 3.1
365 wt.%, Mg# > 91, TiO₂ < 1.5 wt.%, CaO < 9.5 wt.% and Na₂O > 4 wt.%, ii) amphiboles from veins
366 and PH4 and PH9 (22% and 62% vein bearing peridotites) patches have 3.1 wt.% < FeO < 6 wt.%,
367 85 < Mg# < 91, TiO₂ > 2 wt.%, CaO > 9.7 wt.% and Na₂O < 4 wt.%, iii) PH23 (hornblendite)
368 amphiboles have FeO > 7 wt.%, Mg# < 85 and TiO₂ > 2.5 wt.% (Figure 3.b, c and Supplementary
369 Material 4).

370

371 4.2.4 Phlogopite

372 In the Pello Hill xenoliths, there are at least two chemical groups of phlogopite. These
373 within patches have concentrations of $\text{TiO}_2 \leq 2$ wt.%, $\text{Cr}_2\text{O}_3 > 2$ wt.%, $\text{Mg\#} > 92$, and $\text{K}_2\text{O} < 9$
374 wt.%, whereas phlogopites in veins have $\text{TiO}_2 > 3$ wt.%, $\text{Cr}_2\text{O}_3 < 1$ wt.%, $\text{Mg\#} < 92$ and $7.2 <$
375 $\text{K}_2\text{O} < 8.6$ wt.% (Figure 3.d, and Supplementary Material 4). The phlogopites in the vein sample
376 PH23 have a very low Mg# (80.6-81.8) compared to other phlogopites in vein from other Pello
377 Hill xenoliths (Mg# > 88) (Figure 3.d).

378

379 4.2.5 Oxides

380 Two distinct types of oxides are present in the Pello Hill xenolith samples: chromite
381 and ilmenite. Chromites are present in the peridotites or in the metasomatic vein edges. They
382 are interstitial or occur as inclusions in phlogopite, amphibole, and less frequently in
383 clinopyroxene (Figure 2.b, c and d). Ilmenites are only interstitial in the metasomatic veins
384 (PH9 sample, Figure 2.h). The concentration of Cr_2O_3 in chromites ranges from 47 to 60 wt.%.
385 Chromites at the rim of the veins have higher titanium (> 2.5 wt.% TiO_2) and lower chromium
386 concentrations (< 50 wt.% Cr_2O_3) than those located farther within the peridotite (0.05-1.9
387 wt.% TiO_2 and 53.2-60.2 wt.% Cr_2O_3). The titanium concentration is higher in the interstitial
388 chromites (0.5 and 1.6 wt.% TiO_2 in PH26 and PH9, respectively) and chromium content is
389 higher in chromites occurring as inclusions (1.5 and 2 wt.% Cr_2O_3 in PH26 and PH9,
390 respectively) (Supplementary Material 4).

391

392 4.3 Microstructures

393 In the vein-free xenolith, some of the largest olivine porphyroclasts are elongated,
394 parallel to the olivine [100] axis. The subgrain boundaries (undulose extinction) are
395 perpendicular to the olivine [100] axis (Supplementary Material 5). The olivine J-index ranges
396 from 4.00 to 4.79 (Supplementary Material 5). These values are in the lower range of the J-
397 index estimated for worldwide natural olivine aggregates (between 3.2 and 27.0) by (Ismail
398 and Mainprice, 1998).

399 The four peridotite samples with veins have a vein orientation oblique or
400 subperpendicular to the olivine [100] axis (Figure 4). There is no preferential orientation of
401 clinopyroxene, amphibole, and phlogopite in veins, even if the amphibole [001] axis of PH9
402 and PH4 seems to be parallel to the olivine [100] axis.

403 For almost all Pello Hill samples, vein sample PH23 and peridotite PH27 with 80% vein
404 excluded, the olivines have a marked orientation with a point pattern for the [100] axis, and a
405 girdle pattern for [010] and [001] axes (Figure 4 and Supplementary Material 5).

406 The aspect ratios of the main phases (olivine, orthopyroxene, clinopyroxene,
407 amphibole, phlogopite) used for velocity computation correspond to the mean aspect ratio,
408 which is the measured mean longest axis divided by the mean shortest axis. For the full
409 aggregate (peridotite or peridotite + vein) velocity computation, the mean aspect ratios

410 calculated from all measured samples are: 1:1.6:1 for olivine, 1.6:1:0.6 for orthopyroxene and
 411 clinopyroxene, 1:1.7:1 for amphibole, 0.6:1:2 for phlogopite, 1:1:1 for magnetite and 1:1:2.8
 412 for ilmenite. For the seismic properties of oriented veins, we used the average aspect ratio for
 413 clinopyroxene, amphibole, and phlogopite (1.7:1:1).

414

415 **5 Seismic properties of mantle xenoliths**

416 **5.1 Mantle peridotites**

417 In harzburgites PH13 and PH27, and dunites PH8 and PH26, the foliation is marked
 418 by olivine elongation, with lineation sub-parallel to [100] axes and a girdle distribution of [010]
 419 axes perpendicular to the foliation. The fast and slow V_p follow the [100] and [010] olivine
 420 axes, respectively, indicating that olivine controls the velocity amplitude and orientation
 421 (Supplementary Material 5 and Table 2). The fast and slow V_{s1} are along [101] and a
 422 combination of [001] and [100] axes, respectively. V_{s2} presents its maximum along [100] and
 423 its minimum for a combination of [001] and [010] axes, respectively. All vein-free samples
 424 have millimetric olivine with sub-grains. The fast velocities (V_p , V_{s2}) are parallel to the olivine
 425 porphyroclast elongation (Supplementary Material 5, unbroken lines on V_p plots).

426 Except for the dunite PH8 which exhibits lower velocities ($V_p=6.9-7.7 \text{ km.s}^{-1}$ and V_{s1}
 427 $=4.1-4.3 \text{ km.s}^{-1}$, and $V_{s2} = 4.0$ to 4.1 km.s^{-1}), all vein-free samples and samples with low vein
 428 proportions (<22%) have homogeneous velocities with $V_{p_{\min}}$ ranging from 7.3 to 7.5 km.s^{-1} ,
 429 $V_{p_{\max}}$ from 8.4 to 8.5 km.s^{-1} , $V_{s1_{\min}}$ from 4.3 to 4.4 km.s^{-1} , $V_{s1_{\max}}$ from 4.5 to 4.6 km.s^{-1} , $V_{s2_{\min}}$
 430 from 4.1 to 4.2 km.s^{-1} , and $V_{s2_{\max}}$ from 4.3 to 4.4 km.s^{-1} (Table 2). The lower velocities and
 431 maximum V_p/V_s ratio for PH8 are most probably due to the presence of interstitial
 432 phlogopite/amphibole (7.8%).

433

434 **5.2 Clinopyroxene, phlogopite, amphibole-bearing mantle veins**

435 In all clinopyroxene, phlogopite, amphibole-bearing mantle veins, the fast and slow
 436 velocity directions follow the amphibole axes ([001] and [100] axes, respectively). However,
 437 when the proportion of phlogopite is higher than amphibole, the phlogopite also partly
 438 controls the fast ([100] and [010] axes) and the slow ([001] axis) velocities (Supplementary
 439 Material 6).

440 For all mineral assemblage (clinopyroxene, amphibole, and phlogopite), the
 441 computed velocities of the veins at 2 GPa and 1200°C are similar and vary from 3.3 to 5.5 km.s^{-1}
 442 for V_p , 2.4 to 3.1 km.s^{-1} for V_{s1} , and 2.2 to 2.7 km.s^{-1} for V_{s2} (Supplementary Material 6). It
 443 should be noted that there is no clear correlation between velocities and relative phase
 444 proportion of amphibole, phlogopite, and clinopyroxene. The velocity of veins is always lower
 445 than the velocities of peridotites.

446

447 5.3 Mantle peridotites and veins

448 For the peridotite containing a low vein proportion (PH18, 6% of vein), the presence
 449 of the vein in the peridotite decreases the mean velocity from 8.4-7.4 to 8.2-7.1 km.s⁻¹ for V_p,
 450 and from 4.5-4.1 to 4.4-4.0 km.s⁻¹ for V_s. This corresponds to a decrease by 2-4% V_p and 2-
 451 3% V_s compared to the peridotite (PH18, Figure 4 and Table 2). However, the velocity
 452 orientation is still controlled by the olivine orientation (Figure 4).

453 In peridotite containing 22% of veins (with 17-18% of phlogopite/amphibole), the
 454 presence of the veins causes a mean decrease of the velocity from 8.5-7.4 to 7.6-6.5 km.s⁻¹ for
 455 V_p and from 4.6-4.1 to 4.1-3.7 km.s⁻¹ for V_s. That corresponds to a decrease by 10-12% V_p and
 456 9-10% V_s compared to the peridotite (PH9, Figure 4 and Table 2). This vein proportion initiates
 457 a shift of the velocity directions, and the slowest V_p direction does not follow the olivine [010]
 458 axis. The veins and peridotite + vein minimum velocities almost have the same orientation
 459 suggesting that the slowest velocities are controlled by the vein velocity tensor
 460 (Supplementary Material 6 and Figure 4, respectively) and the vein orientation.

461 Compared to the peridotite, peridotites with 62% and 80% of veins (PH4 and PH27)
 462 generate a mean decrease in the velocity from 7.5-7.2 to 4.5-4.6 km.s⁻¹ and from 8.4-8.3 to
 463 6.0-5.4 km.s⁻¹ for V_{p_{min}} and V_{p_{max}} (-38 to -40% and -29 to -35%, respectively), from 4.4-4.1 to
 464 2.9-3.1 km.s⁻¹ and from 4.6-4.3 to 3.0-3.4 km.s⁻¹ for V_{s_{min}} and V_{s_{max}} (-29 to -31% and -25 to -
 465 31%, respectively) (Figure 4 and Table 2). With 60% of phlogopite/amphibole vein assemblage,
 466 the velocity direction completely shifts away from the main-driving-velocity olivine axes
 467 (Figure 4, PH4).

468 In the vein-rich samples, the maximum and minimum V_p of the full sample follow the
 469 orientations of amphibole ([001] and [100] axes respectively) and clinopyroxene ([001] and
 470 [010] axes, respectively) (Figure 4 and Supplementary Material 6). The V_p fast and slow axes
 471 of phlogopite ([010], [100] and [001]) are overwritten by the slow and fast axes of the
 472 amphibole, respectively. The fast V_{s₁} mostly follows an intermediate direction between the
 473 amphibole ([001]) and the phlogopite ([100]) axes. The slow V_{s₁} is oriented parallel to the slow
 474 V_p, while the fast V_{s₁} is uniformly oriented perpendicular to the slow V_{s₁}. The fast V_{s₂} is
 475 oriented parallel to the fast V_p, while the slow V_{s₂} is uniformly oriented perpendicular to the
 476 fast V_p (Supplementary Material 6).

477 The main direction of the velocity also depends on the vein orientation (Kim et al.,
 478 2020 and Supplementary Material 7). The maximum of V_p and V_{s₂} velocities are perpendicular
 479 (PH4, PH27) or oblique/sub-perpendicular to the orientation of the veins (PH18, PH9, Figure
 480 4). The vein orientation follows the main orientation of the olivine [010] axis for a vein
 481 proportion < 60%.

482

483 5.4 Fluid-veins in mantle peridotite

484 The composition of lavas from the NTD ranges from H₂O-CO₂-rich sub-alkaline to
 485 highly alkaline magmas attesting of complex fluid/melt compositions within the lithospheric
 486 mantle (Baudouin and Parat, 2020; De Moor et al., 2013; Lee and Rudnick, 1999; Lee et al.,
 487 2017; Mana et al., 2015, 2012; Muirhead et al., 2020). To account for the variable geochemical

488 composition of magmas and fluid percolation in the mantle, we determined the seismic
489 properties of the mantle for vein-bearing peridotite considering different vein orientations
490 relative to the olivine CPO, and 4 different melt compositions. These melts are: a dry
491 peridotitic melt, used as a reference for dry basalt, a CaCO_3 fluid, which refers to carbonatitic
492 melt, a hydrous peridotitic melt and a CO_2 -bearing peridotitic melt (Table 4; Genge et al., 1995;
493 Sakamaki et al., 2011, 2010, 2009).

494 First, we use AnisEulerSC to calculate the elastic tensor of the peridotite at 2 GPa and
495 1200°C, with mineral proportion, CPO and aspect ratio, as performed for crystallized vein-
496 bearing peridotites (step 1). Then, the tensor of the whole aggregate with fluid-filled
497 inclusions (high pressure and high temperature) is computed from the peridotite elastic tensor
498 and an isotropic fluid tensor. The fluid tensor is defined by the bulk modulus of the fluid (Table
499 4; Genge et al., 1995; Sakamaki et al., 2011, 2010, 2009). We fix the aspect ratio of the
500 inclusions 5:1:5 (lens shape) to model an intermediate geometry between an isolated melt
501 pocket and interconnected pores filled with melt (intergranular melt; Clark and Lesher, 2017).
502 These are the usual shapes of the first liquids observed in partial melting experiments
503 (Hammond and Humphreys, 2000).

504 The seismic properties of PH4 and PH27 cannot be fully modeled with fluid-filled
505 inclusions because the vein proportion is too large and prevents the Vs from propagating
506 (calculated Vs < 0.2 km.s⁻¹). In these samples, it induces unrealistic Vp/Vs (above 18) and P-
507 waves velocities are reduced to less than 3 km.s⁻¹.

508

509 **5.4.1 Orientation of the fluid-filled inclusions in peridotite**

510 Three lens-shape inclusion orientations have been chosen relative to the structural
511 frame (X,Y,Z) of the host peridotite PH9 (Figure 5 and Supplementary Material 7), defined
512 following the observation of the foliation plane (X,Y) and lineation (X). The fastest velocities
513 have orientations parallel to directions that combine the longest axes of the inclusions (A1)
514 and the maximum velocity of the host rock, i.e., parallel to the [100] axis of olivine. For
515 instance, peridotite with 5:5:1 oriented inclusions displays higher maximum velocity than the
516 peridotite with 1:5:5 orientations, where the long axis of the inclusions is perpendicular to the
517 [100] axis of olivine.

518

519 **5.4.2 Velocities for different fluid compositions in lens shape inclusions**

520 For lens-shaped fluid-filled inclusions, the velocities and Vp/Vs ratios vary as a
521 function of fluid composition and aspect ratio for H_2O and CaCO_3 composition (Kendall, 2000;
522 Mainprice, 1997 and Table 3). The velocities and Vp/Vs ratios for CO_2 -bearing peridotitic melt
523 (2.5% of CO_2) and anhydrous peridotitic melt are very similar, with velocity variations lower
524 than 0.05 km.s⁻¹.

525 For CaCO_3 melt and 6% of lens-shape inclusions (PH18), only Vp has a significant
526 change compared to the anhydrous inclusions (+0.09, +0.23 km.s⁻¹ compared to dry melt,
527 equivalent to +1% and +3.4% respectively, Table 3). When the inclusion proportion in the

528 sample is 22% (PH9), the velocities variations do not have the same trend: $V_{p_{\max}}$ increases by
529 0.07 km.s^{-1} (+1%) while $V_{p_{\min}}$ decreases by 0.12 km.s^{-1} (-2%) compared to dry melt.

530 For hydrous lens-shape inclusions (5 wt.% water in the melt), there is an important
531 decrease of V_p compared to anhydrous melt: down to -0.32 km.s^{-1} (equivalent to -5%) when
532 6% of inclusions is present in the sample, and down to -0.85 km.s^{-1} (equivalent to -16%) when
533 there is 22% of inclusions (Table 3). The maximum velocity of V_{s_2} is also significantly impacted
534 with a decrease of -0.16 km.s^{-1} (-5% of variation) with 22% of inclusions.

535

536 6 Discussion

537 6.1 Depth of fluid/melt-rock interaction and geotherm beneath the NTD

538 Characterizing the deep asthenospheric mantle source of alkaline magmas is crucial
539 because it represents the source of the metasomatic agent leading to the formation of veins
540 within the lithospheric mantle. In the southern part of the NTD, the Labait and Pello Hill
541 volcanoes erupted CO_2 -rich nephelinites and melilitites with very close trace element
542 composition (Baudouin and Parat, 2020; Dawson and Smith, 1988) suggesting the presence of
543 a deep phlogopite-carbonate-garnet-bearing source beneath the Lithosphere Asthenosphere
544 Boundary (LAB > 120 km, Baudouin and Parat, 2020; Lee and Rudnick, 1999). Labait volcano
545 carried garnet-bearing xenoliths, whereas at Pello Hill volcano, only spinel-peridotite xenoliths
546 have been sampled. This either suggests a shallower depth of partial melting for highly alkaline
547 parental magmas as observed for the Natron-Engaruka volcanoes (75-90 km, Mattsson et al.,
548 2013), or that melilitite magma did not sample garnet-bearing xenoliths because of different
549 mantle rheology or lower volume of magmas. The few experimental studies carried out at
550 lithospheric conditions (up to 3 GPa, e.g. Condamine et al., 2016) do not allow to estimate the
551 depth of partial melting of peridotite that leads to the formation of alkali magmas. Only one
552 empirical equation has been calibrated for partial melting producing basaltic melts, based on
553 the SiO_2 and MgO contents of the primary melts (Albarède, 1992). Using this empirical
554 equation and the composition of Pello Hill melilitite ($\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O} = 16.34 \text{ wt.}\%$, $\text{SiO}_2+\text{Al}_2\text{O}_3$
555 $= 47.61 \text{ wt.}\%$, Supplementary Material 9), we propose a melting pressure of 5.3 GPa (175 km).
556 This depth is located below the estimated LAB beneath the NTD from surface wave
557 tomography (Craig et al., 2011; Fishwick, 2010; Priestley et al., 2008), and is deeper than the
558 CO_2 -free partial melting experiments (e.g., Adam, 1990; Condamine et al., 2016; Foley et al.,
559 2012; Frey et al., 1978; Green and Falloon, 1998). It corroborates however the deep origin of
560 magma and mantle xenoliths beneath the NTD (this study, and Lee and Rudnick, 1999).

561 The depth and temperature we estimated from mineral equilibrium in the vein-free
562 peridotites from Pello Hill suggest that xenoliths originate from $\sim 37 \text{ km}$ (PH4, dunite) down
563 to 150 km depth (PH13, harzburgite), with temperatures ranging from 1040 to 1200°C (Figure
564 6, Supplementary Material 8). Spinel-phlogopite-bearing Pello Hill Iherzolites (PH17, PH26)
565 originate from 82 to 91 km depth (Figure 6). In contrast, mineral equilibrium in the vein-free-
566 and spinel-bearing harzburgite (PH13, 150 km) indicates deeper mantle origin although garnet
567 was not observed in the xenoliths. The P-T conditions agree with the stability field
568 experimentally determined for phlogopite, extended for 1100-1350°C and 3-6 GPa (90-180
569 km) (Condamine et al., 2016 and Condamine personal communication). The P-T conditions are

570 close to or just above the stability field of spinel in CMAS or MAS experiments (up to 180 km
571 in the harzburgitic mantle, e.g., Klemme, 2004; Ziberna et al., 2013). However, it should be
572 noted that the presence of phlogopite and carbon may influence the experimental barometer
573 calibration determined by Brey and Köhler (1990) and Köhler and Brey (1990) for garnet
574 lherzolite.

575 It was not possible to estimate the depth for most vein-bearing xenoliths because
576 either (i) there is no orthopyroxene in the samples (PH23) or (ii) the chemical equilibrium
577 between orthopyroxene and clinopyroxene or olivine and clinopyroxene was not reached
578 (PH1, PH23).

579 The NTD deep mantle xenoliths (> 100 km) display estimated temperatures
580 consistent with the plume-modified geotherm (Figure 6) (Gibson and McKenzie, 2023; Selway
581 et al., 2014). This is in accordance with several geophysical and geochemical studies, and the
582 presence of a mantle plume upwelling below the Tanzanian Craton and the NTD (Clutier et al.,
583 2021; Huerta et al., 2009; MacDonald et al., 2001; Mulibo and Nyblade, 2013a; Pik et al., 2006;
584 Tiberi et al., 2019; Weeraratne et al., 2003). Shallower lithospheric xenoliths (< 100 km) exhibit
585 higher temperatures than the plume-modified geotherm, ranging from 1040°C to 1200°C
586 (Figure 6). The overall P-T distribution of NTD mantle xenoliths cannot be simply explained by
587 Cenozoic, Proterozoic, cratonic or even mantle plume-modified geotherms (Artemieva, 2009;
588 Selway et al., 2014). The Pello Hill mantle xenoliths rather follow a 1100°C isotherm, near the
589 carbonated peridotite solidus (Dasgupta et al., 2013), and most likely relates to the volatile-
590 bearing peridotite solidus and crystallization temperatures of phlogopite and amphibole. This
591 is in accordance with the presence of CO₂-rich silicate magmas and carbonatites in the region
592 (Dawson and Smith, 1988; De Moor et al., 2013; Muirhead et al., 2020).

593

594 **6.2 Lithospheric mantle geochemical evolution**

595 The presence of veins in mantle xenoliths attests to extensive metasomatism
596 beneath the NTD. Both modal abundances and chemical signatures strongly suggest the
597 occurrence of one or several metasomatic event(s) (Aulbach et al., 2011; Chesley et al., 1999;
598 Cohen et al., 1984; Koornneef et al., 2009; Lee and Rudnick, 1999; Rudnick et al., 1994). In the
599 southern part of the NTD (Labait volcano), few veins have been observed in mantle peridotites
600 (Koornneef et al., 2009; Vauchez et al., 2005) and only a few spinel-peridotites display
601 interstitial phlogopite, suggesting weak modal metasomatism. On the contrary, in the
602 northern part, the Pello Hill xenoliths provide evidence of intense metasomatic events at
603 relatively shallow depths (40-90 km) (Aulbach et al., 2011; Koornneef et al., 2009; this study).

604 The mineral assemblages and composition of veins in the Pello Hill xenoliths (e.g., Fe
605 and Ti-rich ilmenite, low-Cr₂O₃ and high-TiO₂ diopside, phlogopite and amphibole, this study)
606 indicate the presence of FeO- and TiO₂-rich alkaline mafic silicate melts/fluids within the
607 lithospheric mantle. Those melts/fluids are also CaO-rich, as evident from the high CaO
608 concentrations in clinopyroxenes and amphiboles from veins or vein-bearing xenoliths (Figure
609 3). In addition, amphiboles in the veins are potassium-rich, which may be interpreted as the
610 imprint of a subduction-related fluid (e.g., Soltanmohammadi et al., 2018).

611 We also observed chemical variations between the core and rim of minerals, and
 612 depending on the distance from the vein. Orthopyroxenes are zoned with higher Al_2O_3 and
 613 Cr_2O_3 concentrations in the core than in the rim indicating that the enstatite either re-
 614 equilibrated at the rim with a circulating fluid or melt, or that the core of minerals crystallized
 615 deeper than the rim (Longhi et al., 1993). The decrease of Cr_2O_3 concentration observed in
 616 clinopyroxenes of PH4 dunite from the core (2.4-3 wt.%) to rim (1.6-2.1 wt.%) also argues for
 617 metasomatic fluid/melt circulation and re-equilibration of the minerals. In addition, the
 618 chemical gradient observed for clinopyroxene and olivine from the vein to the peridotite may
 619 result from a diffuse metasomatic percolation within the whole rock mantle peridotite
 620 associated to the metasomatic veins (Bodinier et al., 1990). Furthermore, a higher proportion
 621 of veins may indicate a higher volume of melt within the peridotite as well as a longer time of
 622 percolation and re-equilibration between host-peridotite and alkaline melt (e.g., low NiO in
 623 olivine for vein proportion > 50%, PH4).

624 The presence of harzburgites with no or rare phlogopites and amphiboles (PH1 and
 625 PH2, Table 1) indicates that such a refractory lithospheric mantle below Pello Hill underwent
 626 melt extraction, and may be linked to subduction magmatism (Archean, Proterozoic?) or
 627 recent intraplate magmatism (Aulbach et al., 2011; Rudnick et al., 1994). On the other hand,
 628 harzburgites with interstitial phlogopites or amphibole patches (PH13, PH18 and PH24), and
 629 2-10% clinopyroxene-rich PH8 and PH26 dunites, suggest later re-fertilization processes
 630 associated with metasomatic events. The interstitial metasomatic phlogopites and
 631 amphiboles are often associated with chromite with no coronitic reaction, suggesting that the
 632 presence of "primary" spinel may favor phlogopite and/or amphibole growth during
 633 metasomatic event. These patches of minerals are MgO , Cr_2O_3 and Na_2O -rich compared to
 634 those occurring within the veins. These new minerals were formed during a metasomatic
 635 event involving hydrous and potassic fluids/melts. These fluids may be subduction-related
 636 before the Mozambique Belt orogen (Aulbach et al., 2011; Rudnick et al., 1994) or may be
 637 more recent and plume-related, with intermediary composition between carbonatitic and
 638 silicic (Aulbach et al., 2011; Bell and Simonetti, 1996; Mollet et al., 2009). Further studies,
 639 especially isotopic dating, would help to decipher between the metasomatic event(s) and
 640 long-term re-equilibration below Pello Hill (e.g., Koornneef et al., 2009).

641

642 **6.3 Seismic properties of metasomatized lithospheric mantle**

643 **6.3.1 Seismic properties of crystallized-vein lithospheric mantle**

644 In the Pello Hill mantle peridotites, the foliation is marked by the elongation of olivine
 645 porphyroclasts parallel to the [100] axis, and the veins crosscut the foliation (perpendicular to
 646 oblique to the olivine [100] axis) (Figure 4 and Baptiste et al., 2015). The fast and slow V_p
 647 follow the [100] and [010] olivine axes, respectively, indicating that olivine controls the
 648 velocity amplitude and orientation (the fast velocity V_p being parallel to [100] axis). The
 649 presence of weak modal metasomatism (8% interstitial phlogopite and amphibole, PH8)
 650 decreases V_p by about 7 to 9%, V_{s1} by 6%, and V_{s2} by 4 to 5% compared to vein-free
 651 peridotites (Figure 4 and Table 2). Still, it does not modify the velocity orientation, that is
 652 controlled by the olivine orientation (CPO).

653 The computed velocities for vein-bearing peridotites clearly indicate that the
654 proportion of metasomatic minerals (clinopyroxene, amphibole, and phlogopite) decreases
655 V_p and V_s , with phlogopite being the most effective mineral, followed by amphibole and then
656 clinopyroxene (Figure 4, Table 1 and Table 2), in agreement with single crystal seismic
657 properties (Almqvist and Mainprice, 2017). The presence of 22% of crystallized veins in the
658 peridotite decreases V_p by 10 to 12%, and V_{s_1} , and V_{s_2} by 9 to 10% (PH9, Figure 4). These
659 values are close to those previously determined for vein-bearing peridotite xenoliths from
660 Eledoi volcano using MTEX program (not considering grain volume) at ambient pressure and
661 temperature (28% of vein, Baptiste et al., 2015). However, the consideration of the grain
662 volume in AniseulerSC decreases the maximum velocities by 0.6 km.s^{-1} for V_p and V_{s_2} (-7%
663 and -12%, respectively) and 0.5 km.s^{-1} for V_{s_1} and (-12%) compared to Baptiste et al. (2015).
664 Importantly, when the vein-bearing peridotites is with $\geq 60\%$ of vein, the P-wave velocity
665 anisotropy is dominated by the velocity orientation of the vein, and no more by the main
666 olivine orientation (Figure 4).

667

668 6.3.2 Seismic properties of deep fluid/melt in the lithospheric mantle

669 The presence of fluid/melt in the lithospheric mantle beneath the NTD is strongly
670 supported by the presence of recently erupted magmas at the surface (Mana et al., 2015),
671 and by low velocity zones in tomographic images (Mulibo and Nyblade, 2013b; Tiberi et al.,
672 2019). Melt percolation is a deep (175 km, Pello Hill melilitite, Figure 6) and continuous
673 process leading to mantle-rock interactions and changes of mantle rheology. The volume,
674 distribution, and composition of melt may have important implications for the seismic
675 properties of the mantle (Clark and Lesher, 2017). In addition to melt fractions and aspect
676 ratio (Clark and Lesher, 2017; Kendall, 2000; Kendall et al., 2006), the mineralogical variations
677 (Stixrude and Lithgow-Bertelloni, 2005), the presence of volatile species (Karato and Jung,
678 1998), and the anelasticity enhanced by high temperatures (Faul and Jackson, 2005) may
679 contribute to lower seismic velocities.

680 The percolation of deep highly alkaline and alkaline magmas in the lithospheric
681 mantle beneath the NTD is supported by the occurrence of melilitites and nephelinites,
682 whereas the presence of carbonatite and CO_2 -rich and/or H_2O -rich melt inclusions in minerals
683 and hydrous minerals in magmas suggests a wide range of magma compositions (Baudouin et
684 al., 2018, 2016; Baudouin and Parat, 2020; Dawson, 2012; Guzmics et al., 2012).

685 Our computation of seismic properties of natural mantle xenoliths from Pello Hill
686 clearly indicates that the presence of melt and/or fluids in peridotitic rocks decreases both V_p
687 and V_s (Figure 4 vs Figure 5 and Table 2 vs Table 3). The presence of anhydrous peridotitic
688 melt in the crystallized peridotitic mantle would lower the P-velocity by 5 to 8% and S-velocity
689 by 4 to 10% for 6% of the melt fraction (peridotite + fluid/melt velocities in Table 3 and
690 peridotite velocity in Figure 4). This agrees with the elastic properties of silicate melts which
691 predict a P and S wave velocity decrease of up to 9 and 11%, respectively, for 1-5% of the
692 anhydrous melt between 70 and 150 km deep (Clark and Lesher, 2017).

693 Interestingly, the addition of 2.5 wt.% of CO_2 in anhydrous melt does not change the
694 seismic properties compared to anhydrous peridotitic melt (Table 3). However, the lack of

695 experimental determination of bulk modulus for higher CO₂ content at high pressure does not
 696 allow evaluation of carbonate-rich composition effect on seismic properties (CO₂ solubility in
 697 alkaline magmas can be as high as 15%, Brooker et al., 2001; Moussallam et al., 2015).
 698 Nevertheless, the presence of 5 wt.% of water in 22% of peridotitic melt in the crystallized
 699 peridotitic mantle strongly reduces the seismic velocities down to 39% for $V_{p_{\min}}$ and down to
 700 33-34% for $V_{s_{1\min}}$ and $V_{s_{2\min}}$ (Table 3 and Figure 4). It should be noted that when fluid
 701 inclusions are added in the crystallized peridotitic mantle, the S velocities decrease faster than
 702 the P velocities, except from hydrous melt. This is due to the low bulk modulus used in the
 703 computation which accounts for the supercritical behavior of water. This means that at the
 704 depth/temperature conditions at which we modeled the xenoliths, water behaves as a gas,
 705 leading to a faster decrease of V_p than V_s .

706 Numerous experimental and physical modeling (Clark and Lesher, 2017 and
 707 references herein) demonstrate that the distribution of the melt in the peridotite may have a
 708 strong influence on seismic velocity. In the case of computed fluid-filled inclusions in mantle
 709 xenoliths, the fast velocities have orientations parallel or sub-parallel to the longest axes of
 710 the inclusions (A1). This is observed for all inclusion orientations regardless of the fluid's
 711 nature/composition (Figure 5). The maximum velocity is then obtained when the CPO of
 712 olivine ([100] axis orientation) and maximum V_p and V_{s_2} are parallel to the longest axis of the
 713 inclusion (parallel to A1) (Figure 4 and Figure 5).

714 For a peridotitic melt (anhydrous, 5% hydrous or with 2.5% CO₂), the lower bound
 715 velocities values for lens-shape inclusions are lower and the V_{s_1} upper bound values are higher
 716 than for acicular inclusions. Two patterns are competing: seismic components perpendicular
 717 to the longest lens axes (perpendicular to A1 and A2) have to bypass the inclusion, whereas
 718 seismic components parallel to the longest axes are "guided" along the inclusion elongation
 719 (Figure 5).

720 Metasomatism can be viewed as present-day ascent of volatile-rich small-fraction
 721 melts or past crystallized veins. The superposition of both veins would have a strong effect on
 722 seismic velocities. However, if we assume that crystallized veins with clinopyroxene-
 723 amphibole-phlogopite have a composition close to a peridotitic melt with water, the presence
 724 of melt would further reduce velocities compared to crystallized veins. For peridotite with 6%
 725 of lens-shaped inclusions filled with peridotitic melt with 5% H₂O, V_p decreases by 4 to 9% and
 726 V_s decreases by 1 to 7% compared to peridotite with crystallized veins (PH18). For peridotite
 727 with 22% lens-shaped inclusions filled with peridotitic melt with 5% H₂O (PH9), V_p decreases
 728 by 9 to 31% and V_s decreases by 7 to 28% (Figure 4 vs Figure 5).

729

730 **6.4 In situ seismic properties versus travel-time tomography**

731 The scale over which petrophysics and seismic tomography consider the seismic wave
 732 propagation varies from tens of centimeters up to several hundreds of kilometers,
 733 respectively. At the kilometer-scale, the travel-time tomography reflects a current regional
 734 and integrative effect of temperature, composition and anisotropy, whereas the seismic
 735 properties calculations on xenoliths only refer to local properties for a given composition at a
 736 specific depth (2 GPa in this study).

737 We propose here to use our in situ seismic property results to explain recent
 738 tomographic images in the NTD (Clutier et al., 2021). To transfer our natural sample-scale
 739 results to the regional scale, we first have to convert the mantle xenolith seismic variations
 740 into tomographic variations (Clutier et al., 2021). We compute the percentage difference
 741 between the IASP91 model (Kennett and Engdahl, 1991) and the mantle xenolith-estimated
 742 velocities as follows:

$$743 \quad \Delta V = \frac{V_x - V_i}{V_i} \times 100 \quad (3)$$

744 where V_x is the mantle xenoliths estimated V_p , and V_i is the velocity of IASP91 model (Kennett
 745 and Engdahl, 1991), also used as the reference in the P-wave tomography (Figure 7), at the
 746 mantle xenolith depth (40-90 km). The velocity variations relative to the IASP91 model (ΔV)
 747 are reported in Table 2 and Table 3. It should be noted that in relative arrival-time
 748 tomography, because the regional model is lost during the process, the 0% value of the final
 749 tomographic model may be different from the real global average (Bastow, 2012). In Clutier
 750 et al. (2021) tomographic inversion, the insertion of an a priori 3D crustal velocity model
 751 restricts this problem of shift between the regional mean model and the IASP91 mean model.
 752 Therefore, we can compare the velocity variations from the P-wave tomography and the ones
 753 from the xenoliths computed with equation (3). Moreover, to check this latter statement, we
 754 calculated the difference between the mean absolute velocity of the Clutier et al. (2021) in
 755 the resolved area, and the value of IASP91 at 80, 135 and 185 km depth. The Clutier et al.
 756 (2021) model has a maximum variation of -1.02% compared to IASP91 (-0.49% at 80km, -
 757 0.46% at 135km and -1.02% at 185km).

758 The travel-time tomography beneath the NTD shows that a low-velocity anomaly is
 759 present beneath the rift axis from the Moho (35 km) down to the asthenosphere (Figure 7 and
 760 Clutier et al., 2021). This central negative anomaly is flanked by two high-velocity anomalies,
 761 on the west side beneath the Tanzanian Craton and on the east side beneath the Proterozoic
 762 belt. The P velocity variations from tomography range from -15% to +9% for the whole area.
 763 It is restricted between -15% and -9% for the xenolith estimated location beneath Pello Hill
 764 (depth range from 40 to 90 km, Figure 7).

765 Such a range in V_p anomalies, also reported in density anomalies (Tiberi et al., 2019),
 766 cannot be accounted for by only a temperature effect. First, the depth of the partial melting
 767 estimated from Pello Hill melilitite is located at the downward border of the low-velocity
 768 tomographic anomaly at -2% of P velocity variation (Figure 7). Then, an increase of 100°C
 769 would decrease V_p by ca. 0.8% at 100 km and for a 1100-1200°C isotherm (Cammarano et al.,
 770 2003). The temperatures estimated from olivine-pyroxene equilibrium in mantle xenoliths
 771 suggest a temperature range isotherm of 1040 to 1200°C beneath the rift axis. These
 772 temperatures are then 300°C (80-90 km depth) and 600°C (40 km depth) higher than the
 773 plume-modified cratonic geotherm (Figure 6). Following this linear decrease of V_p with
 774 increasing temperature, such temperature variations can account for only 2.4% anomalies at
 775 80-90 km depth and 4.8% close to the Moho. Even if anelasticity may introduce non-linearity
 776 into temperature-seismic velocity relationship that are not taken into account here
 777 (Cammarano et al., 2003; Goes et al., 2000; Karato, 1993; Kuskov and Kronrod, 2006; Sobolev
 778 et al., 1996), the presence of metasomatic veins is probably the most robust candidate to
 779 explain velocity anomaly values of \sim -10% within the lithosphere.

780 To compare the seismic properties of crystallized vein-bearing peridotites and IASP91
781 model, we calculate the effect of vein proportion on $V_{p_{\max}}$ and $V_{p_{\min}}$ for the vein bearing
782 samples (PH4, PH9, PH18 and PH27), and correct the resulting velocities from temperature
783 effects (1200°C instead of ambient temperature, Supplementary Material 3). We computed
784 the velocities in three steps (see section 3.4.2) using the elastic properties of vein-free
785 peridotite and the ones from veins from the xenoliths PH4, PH9, PH18 and PH27. The relative
786 difference of the maximum and minimum V_p to IASP91 are plotted as a function of crystallized
787 vein proportion in Figure 8.a. It also should be mentioned that for the calculated ΔV_p of
788 peridotites with vein, $V_{p_{\max}}$ is slightly higher or equal to the IASP91 model ($\Delta V_{p_{\max}} = +0.6$ to
789 $+4\%$, Figure 8.a) and $V_{p_{\min}}$ is lower than the IASP91 model ($\Delta V_{p_{\min}} = -5$; -9% , Figure 8.a). This
790 difference is due to local anisotropy, which indicates higher temperatures and the percolation
791 of a partial-melting front across the lithosphere.

792 The decrease of P velocity can be also explained by the presence of lens-shape fluid-
793 filled inclusions (Figure 5 and Table 3). As with the crystallized veins, we calculated the vein
794 proportion effect on the velocities for vein bearing samples at 2 GPa and 1200°C (Figure 8 .b
795 to .d), but in two steps (see section 3.4.2), using the elastic properties of vein-free peridotite
796 (at 2 GPa and 1200°C) and the ones from fluids (Table 4). Lens-shape hydrous melt would
797 more efficiently reduce the P velocity in peridotite. The presence of 6 to 22% of lens-shape
798 hydrous melt would decrease $V_{p_{\text{int}}}$ by 15 to 20% (Table 3 and Figure 8.b to .d, depending on
799 inclusion orientation). Comparatively, for the same proportion of vein in the peridotite, $V_{p_{\text{int}}}$
800 would decrease by 13 to 17% with lens-shape dry or 2.5%-CO₂ melt, and by 9 to 14% with lens-
801 shape CaCO₃ melt (Figure 8.b to .d). Compared to crystallized veins, the mean velocity
802 variations of fluid-filled inclusions are higher for hydrous, anhydrous and 2.5%- CO₂ melts,
803 while they are similar for CaCO₃ melts.

804 Microstructures and CPO analysis of the Tanzanian xenoliths show that the
805 lithospheric mantle in the Northern Tanzanian Divergence zone is pervasively deformed
806 (Baptiste et al., 2015; Vauchez et al., 2005; this study). However, the strong heterogeneity in
807 mantle xenoliths, microstructures, and olivine CPO suggest that this deformation was
808 acquired during multiple tectonic events (Aulbach et al., 2011; Baptiste et al., 2015; Koornneef
809 et al., 2009). Among peridotite from the rift axis, the most common olivine CPO pattern is the
810 axial-[100] pattern, consistent with a strong simple shear component and a transtensional
811 strain regime in the mantle lithosphere beneath the central part of the East-African rift
812 (Baptiste et al., 2015; Tommasi et al., 1999). For these conditions, Vauchez et al. (2005) and
813 Baptiste et al. (2015) suggest that the foliation is expected to be near vertical and the lineation
814 close to horizontal and slightly oblique to the rift axis (Figure 9). Kendall et al. (2006) also
815 suggested that vertical structures, such as vertical melt pockets oriented parallel to the rift
816 axis, are also required in addition to olivine CPO to explain shear wave anisotropy for the more
817 evolved northern part of the East African Rift System.

818 The maximum P velocity ($V_{p_{\max}}$) is representative of velocity parallel to the alignment
819 of the olivine [100] axis, i.e., parallel to the foliation and perpendicular to the crystallized vein
820 for peridotite with low vein proportions (PH18, Figure 4). However, the vein is oblique to the
821 foliation and to the alignment of olivine [100] axis for high crystallized vein proportion (PH4,
822 PH27). For lens-shaped fluid-filled melt inclusions, the maximum P velocity is parallel to the
823 lens inclusion, while the minimum velocity is perpendicular to the maximum elongation of the
824 inclusion.

825 Then, considering the incident tomographic P-wave being vertical, a vertical foliation
826 with horizontal lineation, and the presence of vertical metasomatic veins in the mantle
827 crosscutting obliquely or perpendicularly the foliation, we expect the global P wave path to
828 follow the mean P velocity ($V_{p_{int}}$) for crystallized veins and for fluid-filled veins of the modeled
829 mantle xenoliths velocities (Figure 8, and between models c) and d) in Figure 9). In other
830 words, the P velocity anomalies from tomography (-9 to -15% of V_p variation) could be
831 explained by high-temperature anomalies (200°C at 80-90 km) and the presence of 12 to 25%
832 of crystallized veins oblique to the vertical foliation, or between 8 and 15% of lens-shape fluid-
833 filled melt inclusions with a vertical orientation (Figure 9.d). Any hybrid composition between
834 these two end-members can also fulfill the conditions.

835 The lower bounds of our $V_{p_{min}}$ estimate (6-22% in Table 2, 6% in Table 3, and 6-15%
836 in Figure 8) are just above the mean vein proportion in the lithospheric mantle for MORB
837 source regions, around 2-5%, reported by Hirschmann and Stolper (1996). Our estimates
838 remain a mean/lower bound for vein proportion in NTD lithospheric mantle, as other
839 processes may favor higher values. The low Mg# in olivine in in-rift vein-bearing dunite
840 xenoliths (PH23, and Baptiste et al., 2015) might represent melt channels, or melt
841 accumulation levels within the mantle lithosphere, concentrating vein proportion in specific
842 areas (Berger, 1985; Berger and Vannier, 1984; Kelemen et al., 1995; Kelemen and Dick, 1995;
843 Tommasi et al., 2004). The presence of numerous and undeformed hydrous phases in veins or
844 diffuse pockets in mantle xenoliths suggests extensive post-kinematic metasomatism by
845 hydrous fluids or melts. The presence of deep asthenospheric alkaline magmas located at the
846 downward border of the low-velocity tomographic anomaly at -2% of P velocity variation and
847 close to the plume-modified geotherm (Figure 7 and Figure 6, respectively) may correspond
848 to the presence of deeply isolated melt pockets or interconnected melt (Clark and Leshar,
849 2017; Hammond and Humphreys, 2000). The presence of Archean metasomatic events from
850 ancient subduction zones and the presence of numerous volcanoes at the NTD may contribute
851 to a very high proportion of melt (Fritz et al., 2013; Koornneef et al., 2009; Mulibo and
852 Nyblade, 2013b; Pik et al., 2006; Tiberi et al., 2019). The maximum proportion of crystallized
853 and fluid-filled veins in metasomatic peridotites would, however, be less than 40% and 30%,
854 respectively (Figure 8) when considering $\Delta V_{p_{max}}$ for natural mineral assemblage in veins and
855 peridotitic melt (dry, hydrous and 2.5%-CO₂).

856 The petrological and petrophysical studies of xenolith samples provide constraints to
857 tomographic images in terms of composition and local structure of the mantle. The distinction
858 between the presence of crystallized melt and/or fluid could be further addressed with S-wave
859 lithospheric tomography below the NTD. The relative differences in both amplitude and
860 location of the V_p and V_s anomalies could give us relevant information to identify the presence
861 of melt/fluids and compositional variations (Ito et al., 1979; Watanabe, 1993). V_p is more
862 sensitive to the presence of gas, whereas V_s will drop faster when some melt is present. Then,
863 the tomographic V_p/V_s ratio would be essential to extract more information about the nature
864 of fluids (crystallized or melt-bearing veins) present in the magmatic area of the NTD (Husen
865 et al., 2004; Julian et al., 1998; Nakajima et al., 2001). Coupling the geochemical and
866 petrophysical study of xenolith samples and tomography can provide 3D information on
867 mantle dynamics, while paying attention to the change in spatial and temporal scale, since
868 tomography images the current state of the mantle.

869

870 **7 Conclusions**

871 We combined petrological and seismic property modeling of mantle xenoliths to
872 characterize the impact of metasomatic processes in the lithospheric mantle beneath the NTD
873 on the teleseismic P-wave tomography. First, the presence of an isotherm at 1040-1200°C
874 from olivine-pyroxene equilibria in peridotite beneath the NTD reveals temperature
875 anomalies up to 600°C compared to regional geotherm. This difference can explain 2.4 to 4.8%
876 of Vp decrease at lithospheric depth compared to global earth models. In addition, the seismic
877 properties of vein-bearing and vein-free mantle xenoliths sampled from the in-rift Pello Hill
878 volcano indicate a significant influence of vein proportion, orientation, and composition on
879 seismic properties.

880 When considering vertical foliation with azimuth parallel to the rift axis and
881 horizontal lineation, the presence of 12 to 25% of amphibole-phlogopite-clinopyroxene-
882 bearing crystallized veins can decrease of a vertically incident P-wave by 9 to 15% at 40-90 km
883 depth. The presence of active volcanoes beneath the rift argues for the presence of present-
884 day melt percolation. We estimate the partial melting for alkaline magmas to take place below
885 the LAB. The melt percolation (presence of fluid-filled veins) could explain the observed
886 tomographic low- velocity zone, which amplitudes are mainly controlled by the aspect ratio
887 of the inclusion and the nature of the fluid. 8 to 15% (depending on the melt composition) of
888 vertical lens-shape inclusions filled with melt would be necessary to decrease the tomographic
889 velocities by 9 to 15% at 40-90 km depth.

890

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901

902

903 **Figure captions**

904

905 *Figure 1: Tecto-volcanic map of the NTD. The first-order geological units are indicated (Dawson*
906 *and Smith, 1988; Fritz et al., 2013; Smith and Mosley, 1993): orange for the Tanzanian Craton,*
907 *blue and green for the Western and Eastern parts, respectively, of the Mozambique Belt. The*

908 *inferred edges of the Masai block are represented by the blue dashed line. The major normal*
 909 *faults are represented by the continuous black lines: EBF = Eyasi Border Faults, NBF = Natron*
 910 *Border Fault, MBF = Manyara Border Fault, OOF = Ol Doinyo Ogol Fault, PBF = Pangani Border*
 911 *Fault. Green triangles indicate the location of the main volcanoes. The red star indicates the*
 912 *sample area of this study (Pello Hill). The ages of the volcanism are written in italic next to the*
 913 *volcanic edifices (Dawson, 1992; Macintyre et al., 1974; Mana et al., 2015, 2012; Manega,*
 914 *1993; Mollel, 2007; Nonnotte, 2007; Sherrod et al., 2013; Wilkinson et al., 1986). The two N-S*
 915 *and E-W chemical axes are represented by the green and purple arrows, respectively.*
 916 *Volcanoes abbreviations: B = Burko, El = Eledoi, Em = Embakai, Es = Essimngor, G = Gelai, H =*
 917 *Hanang, Ke = Kerimasi, Ket = Ketumbeine, Ki = Kilimanjaro, Kw = Kwahara, Lab = Labait, La =*
 918 *Lashaine, M = Monduli, Me = Meru, Ng = Ngorongoro, OL = Ol Doinyo Lengai, Olm = Olmani,*
 919 *PH = Pello Hill, T = Tarosero. The East-West red dashed line represents the track of the P-wave*
 920 *tomographic vertical cross-section in Figure 7.*

921 *Figure 2: Photomicrographs of Pello Hill mantle xenolith thin sections in plane polarized light:*
 922 *a) orthopyroxene, with amphibole inclusion, and reaction rim with clinopyroxene (PH13); b)*
 923 *interstitial clinopyroxene, orthopyroxene and chromite in PH18 dunite; c) Typical multi-*
 924 *millimeter amphibole/phlogopite/clinopyroxene patch surrounding chromites in PH13; d)*
 925 *amphibole-clinopyroxene aggregates around chromites in PH4 dunite; e) Diffused vein border*
 926 *with fine amphibole and clinopyroxene grains, and well-crystallized grains within the major*
 927 *vein in PH18; f) Distinct vein/peridotite boundary in PH4 dunite; g) Multi-millimeter phlogopite*
 928 *surrounded by aggregates of clinopyroxene-phlogopite-amphibole-olivine of 100-500 μm in*
 929 *the PH23 vein sample; h) Typical vein composition with clinopyroxene, amphibole, phlogopite,*
 930 *and ilmenite (PH9). Mineral abbreviations: Amph = amphibole; Chr = chromite; Cpx =*
 931 *clinopyroxene; Ilm = Ilmenite; Ol = olivine; Opx = orthopyroxene; Phl = phlogopite. Red*
 932 *continuous and dashed lines delimit the distinct and diffused vein/peridotite boundary,*
 933 *respectively.*

934 *Figure 3: Major element compositions for a) clinopyroxenes, b) and c) amphiboles, and d)*
 935 *phlogopites from the Pello Hill mantle xenoliths. Triangles are for in-vein minerals, circles for*
 936 *in-peridotite minerals and crosses for lava minerals. $\text{Mg\#} = \text{Mg}/(\text{Mg}+\text{Fe}^{2+})\times 100$.*

937 *Figure 4: Comparison between the whole sample, the peridotite (without vein), and the vein*
 938 *seismic properties (each one being computed in one step) for PH18, PH9, PH4, and PH27*
 939 *xenoliths, at pressure 2 GPa and 1200°C. The peridotite was computed directly with*
 940 *AnisEulerSC while the peridotite + vein and the vein were corrected from the temperature*
 941 *effect by applying a percentage decrease (Supplementary Material 3). The proportion of the*
 942 *vein in a sample is provided in brackets next to the sample name. All plots are lower*
 943 *hemisphere. Vein orientations are represented by the red lines on the olivine [100] axis pole*
 944 *figures. The black dashed lines on the V_p plot represent the elongation azimuth of sub-grains*
 945 *(undulose extinction) in the peridotite. N is the number of measurements used to compute the*
 946 *J-index (J) for the olivine.*

947 *Figure 5: Seismic property calculations with different lens-shape inclusion orientations and*
 948 *fluid filling in PH9 22% vein-bearing xenolith sample. Calculations are at pressure 2 GPa and*
 949 *temperature 1200°C with AnisEulerSC. The first row shows, a sketch of the orientation of the*
 950 *inclusions in the sample reference frame, where (X,Y) is the foliation plane and X the lineation*
 951 *(olivine [100] axis). A1 and A3 are the longest and shortest ellipse axes, respectively. A1 = A2*

952 = 5 and $A3 = 1$. AZ stands for the azimuth (counter clockwise) of the ellipse axes and INC for
 953 their vertical inclination. The velocity plots are lower hemisphere, and the maximum and
 954 minimum velocities are marked by a black square and a white circle, respectively.

955 *Figure 6: Calculated temperature and pressure for Pello Hill xenoliths with Brey and Köhler*
 956 *(1990) and Köhler and Brey (1990) thermobarometers (orthopyroxene-clinopyroxene and*
 957 *olivine-clinopyroxene couples). Lashaine P-T conditions are from Rudnick et al. (1994) and*
 958 *Gibson et al. (2013). The Labait xenoliths P-T conditions are from Lee and Rudnick (1999) and*
 959 *Koornneef et al. (2009). The P-T conditions of the partial melting of the Labait nephelinite are*
 960 *calculated from Baudouin and Parat (2020). Additional data of vein-free xenoliths from Pello*
 961 *Hill (black crosses, Dawson and Smith, 1988) were used to calculate P-T conditions as a*
 962 *comparison. The Pello Hill and Labait Moho were determined with receiver functions from*
 963 *Plasman et al. (2017) and A. Clutier personal communication, respectively. The LAB depth*
 964 *below the Labait and Pello Hill were estimated by Craig et al. (2011) with surface wave*
 965 *tomography (vertical resolution of ± 25 km). Cratonic geotherm (44 mW.m^{-2} , Selway et al.,*
 966 *2014), plume modified geotherm (50 mW.m^{-2} , Gibson and McKenzie, 2023), Proterozoic and*
 967 *Cenozoic geotherms (Artemieva, 2009) are also reported. The two peridotite solidi and the*
 968 *modeled melt CO_2 isopleths (thin grey lines, with CO_2 wt.% in the melt) are from Dasgupta et*
 969 *al. (2013).*

970 *Figure 7: West-East vertical cross-section at latitude 2.76°S (cross-section track on Figure 1) in*
 971 *the P-wave tomographic model from Clutier et al. (2021), with the depth of Pello Hill mantle*
 972 *xenoliths (grey triangles and diamonds) and lava partial melting (white star) from this study.*
 973 *Contours of velocity variations are every 3%. The LAB is taken from Craig et al. (2011) (vertical*
 974 *resolution of ± 25 km). Abbreviations: Kit. = Kitumbeine volcano; NBF = Natron Border Fault;*
 975 *OL = Ol Doinyo Lengai; OOF = Ol Doinyo Ogol Fault; PH = Pello Hill volcano.*

976 *Figure 8: Maximum (square), minimum (triangle) and mean (crosses) P-wave velocity*
 977 *variations (ΔV_p) computed for a) the crystallized vein-bearing mantle xenoliths with equation*
 978 *(3) and with temperature correction (1200°C), and b-d) for lens-shaped fluid-filled inclusions*
 979 *in PH9 mantle xenoliths: b) vertical veins perpendicular to the foliation (1:5:5), c) vertical veins*
 980 *parallel to the foliation (5:5:1), d) horizontal veins perpendicular to the foliation (5:1:5). The*
 981 *red and orange straight lines delimit the range of the tomographic low velocity variations*
 982 *induced by the metasomatizing fluid circulation. In graph a), the variation of velocity between*
 983 *the different samples for a common vein percentage is induced by the mineralogical*
 984 *assemblage.*

985 *Figure 9: Simplified structural diagram of a west-east profile in the North Tanzanian*
 986 *Divergence with representations (a-d) of possible relationship between fluid-filled veins (yellow*
 987 *planes) and the peridotites from natural xenoliths in the mantle. The orientation of the seismic*
 988 *velocities in the NTD are constrained by the vertical foliation and horizontal lineation in*
 989 *peridotite (with no veins) far from the rift axis (a), and by foliation, horizontal lineation and*
 990 *veins (perpendicular to the rift) in the rift axis (d), i.e. with intermediate vertical velocity. Spinel-*
 991 *garnet boundary is from Lee et al. (2000). Diagram modified from Kendall et al. (2006).*

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993 **Tables**

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995 *Table 1: Modal composition of Pello Hill mantle xenoliths determined by EBSD.*

Sample	Vein proportion in sample (%)	Total sample modal composition (%)						Vein composition (%)			
		Ol	Opx	Cpx	Phl	Amph	Ox	Cpx	Phl	Amph	
PH2	Harzburgite	0	77	21	0	0	1	0			
PH8	Dunite	0	84	0	7	0	7	0			
PH13	Harzburgite	0	85	13	0	0	1	1			
PH24	Harzburgite	0	85	13	1	0	1	0			
PH26	Dunite	0	92	4	2	1	0	0			
PH18	Dunite	6	85	6	5	2	2	0	45	37	18
PH9	Dunite	22	71	2	8	5	12	2	21	19	60

PH4	Dunite	62	33	0	40	19	8	0	59	29	12
PH27	Harzburgite	80	27	2	16	20	33	3	23	27	50
PH23	Vein	100	16	0	17	20	46	0			

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998 *Table 2: Seismic properties computed with AnisEulerSC at 2 GPa and 1200°C. Velocity variations are relative to the IAPS91 Earth model (Kennett*
 999 *and Engdahl, 1991) for Pello Hill mantle xenolith samples, crystallized full aggregates.*

Sam ple	Vein propo rtion in sampl e (%)	Seismic properties - AnisEuler													Variations relative to IAPS91	
		Vp max	Vp min	A V p	Vs1 max	Vs1 min	AV s1	Vs2 max	Vs2 min	AV s2	Vp/Vs 1 max	Vp/Vs 1 min	Vp/Vs 2 max	Vp/Vs 2 min	ΔVp max (%)	ΔVp min (%)
PH2	0	8.4	7.5	11 .3	4.6	4.4	4. 1	4.4	4.2	4. 8	1.9	1.7	1.9	1.8	4	-7
PH8	0	7.7	6.9	10 .9	4.3	4.1	4. 3	4.1	4.0	4. 0	1.8	1.7	1.9	1.7	-4	-14
PH1 3	0	8.5	7.5	13 .3	4.6	4.4	4. 3	4.4	4.2	6. 0	1.9	1.7	1.9	1.8	6	-7
PH2 4	0	8.4	7.5	11 .6	4.6	4.4	4. 5	4.4	4.2	5. 4	1.9	1.7	1.9	1.8	5	-7
PH2 6	0	8.4	7.3	14 .3	4.5	4.3	4. 2	4.3	4.1	6. 6	1.9	1.7	1.9	1.8	4	-10

PH1 8	6	8.2	7.1	14 .3	4.4	4.3	4. 4	4.3	4.0	6. 2	1.9	1.7	1.9	1.8	2	-11
PH9	22	7.6	6.5	15 .3	4.1	4.0	2. 7	4.0	3.7	6. 9	1.9	1.6	1.9	1.7	-6	-19
PH4	62	6.0	4.5	27 .9	3.4	3.1	10 .5	3.1	2.9	7. 4	1.8	1.4	1.9	1.5	-26	-44
PH2 7	80	5.4	4.6	16 .4	3.2	3.0	4. 9	3.0	2.9	3. 0	1.7	1.5	1.8	1.6	-32	-43
PH2 3	100	4.9	3.7	28 .2	2.8	2.6	7. 8	2.6	2.5	6. 3	1.8	1.4	1.9	1.5	-39	-54

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Table 3: Seismic properties computed with AnisEulerSC at 2 GPa and 1200°C. Velocity variations are relative to the IAPS91 Earth model (Kennett and Engdahl, 1991) for Pello Hill mantle xenolith samples with vertical 5:5:1 fluid-filled inclusions.

Sample	Proportion of vein in sample (%)	Fluid/melt	Seismic properties with fluid or melt-filled veins - AnisEuler														Variations relative to IAPS91	
			Vp max	Vp min	A V p	Vs ₁ max	Vs ₁ min	A Vs ₁	Vs ₂ max	Vs ₂ min	A Vs ₂	AVs max	Vp/Vs ₁ max	Vp/Vs ₁ min	Vp/Vs ₂ max	Vp/Vs ₂ min	ΔVp max (%)	ΔVp min (%)
PH18	6	Dry peridotitic melt	8.0	6.8	15.8	4.3	4.0	8.4	4.1	3.8	8.7	12.5	1.9	1.7	2.0	1.8	-1	-16
PH9	22		7.1	5.4	28.2	3.8	2.9	26.8	3.3	2.7	18.3	30.7	1.9	1.7	2.5	1.8	-11	-33
PH18	6	CaCO ₃ melt	8.1	7.0	13.8	4.4	4.0	8.3	4.1	3.8	9.2	12.4	1.9	1.7	2.0	1.8	0	-13
PH9	22		7.2	5.2	31.6	3.9	3.0	27.1	3.3	2.8	17.5	31.0	1.9	1.7	2.4	1.8	-11	-35

PH1 8	6	Peridotitic melt + 2.5% CO ₂	8.0	6.8	16.0	4.3	4.0	8.4	4.1	3.8	8.7	12.5	1.9	1.7	2.0	1.8	-1	-16
PH9	22		7.2	5.4	28.9	3.8	2.9	26.9	3.3	2.8	18.2	30.7	1.9	1.7	2.5	1.8	-11	-33

PH1 8	6	Peridotitic melt + 5% H ₂ O	7.9	6.5	19.7	4.3	4.0	8.6	4.1	3.8	7.9	12.6	1.9	1.6	2.0	1.7	-2	-20
PH9	22		6.9	4.5	42.3	3.8	2.9	28.2	3.1	2.7	14.2	32.1	1.9	1.6	2.4	1.6	-14	-44

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1006 *Table 4: Bulk moduli used for the computation of velocities for the fluid-filled vein-bearing*
1007 *samples in Figure 5, at 1-4 GPa.*

Fluid	Bulk modulus (GPa)	Density (g/cm³)	Reference
Dry peridotitic melt	24	2.935	Sakamaki <i>et al.</i> 2010
CaCO ₃ melt	18.87	2.35	Genge <i>et al.</i> 1995
Peridotite melt + 5% H ₂ O	8.8	2.75	Sakamaki <i>et al.</i> 2009
Peridotite melt + 2.5% CO ₂	22.9	2.7	Sakamaki <i>et al.</i> 2011

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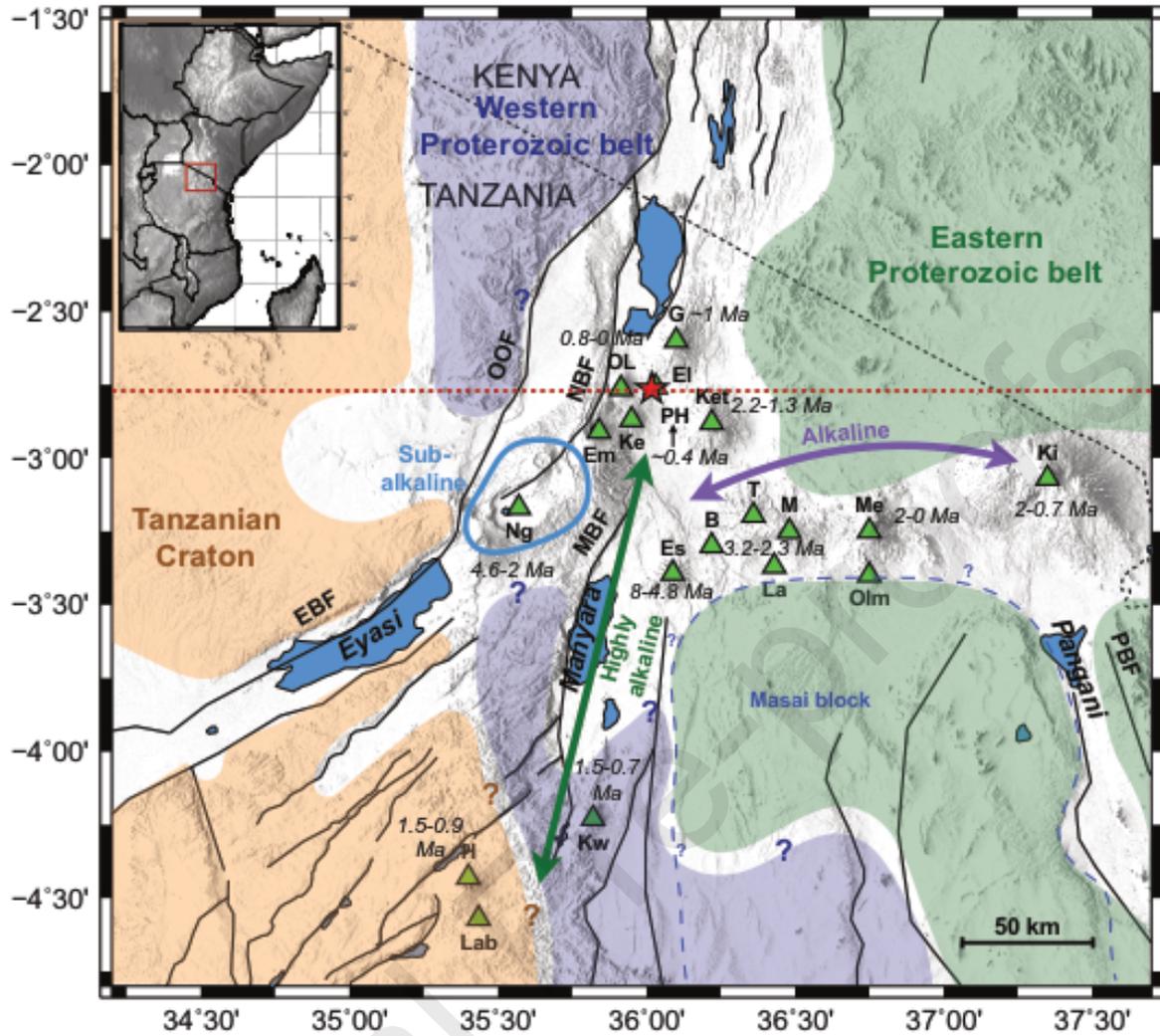
1507 **3 Highlights (<85 characters):**

- 1508 - Mantle at rift initiation is metasomatized by alkaline fluids at 40-90 km depth
- 1509 - 1100°C mantle isotherm relates to the volatile-bearing peridotite solidus
- 1510 - Lithospheric mantle has ≈20% of crystallized or ≈10% of fluid-filled vertical veins

1511

Figure1

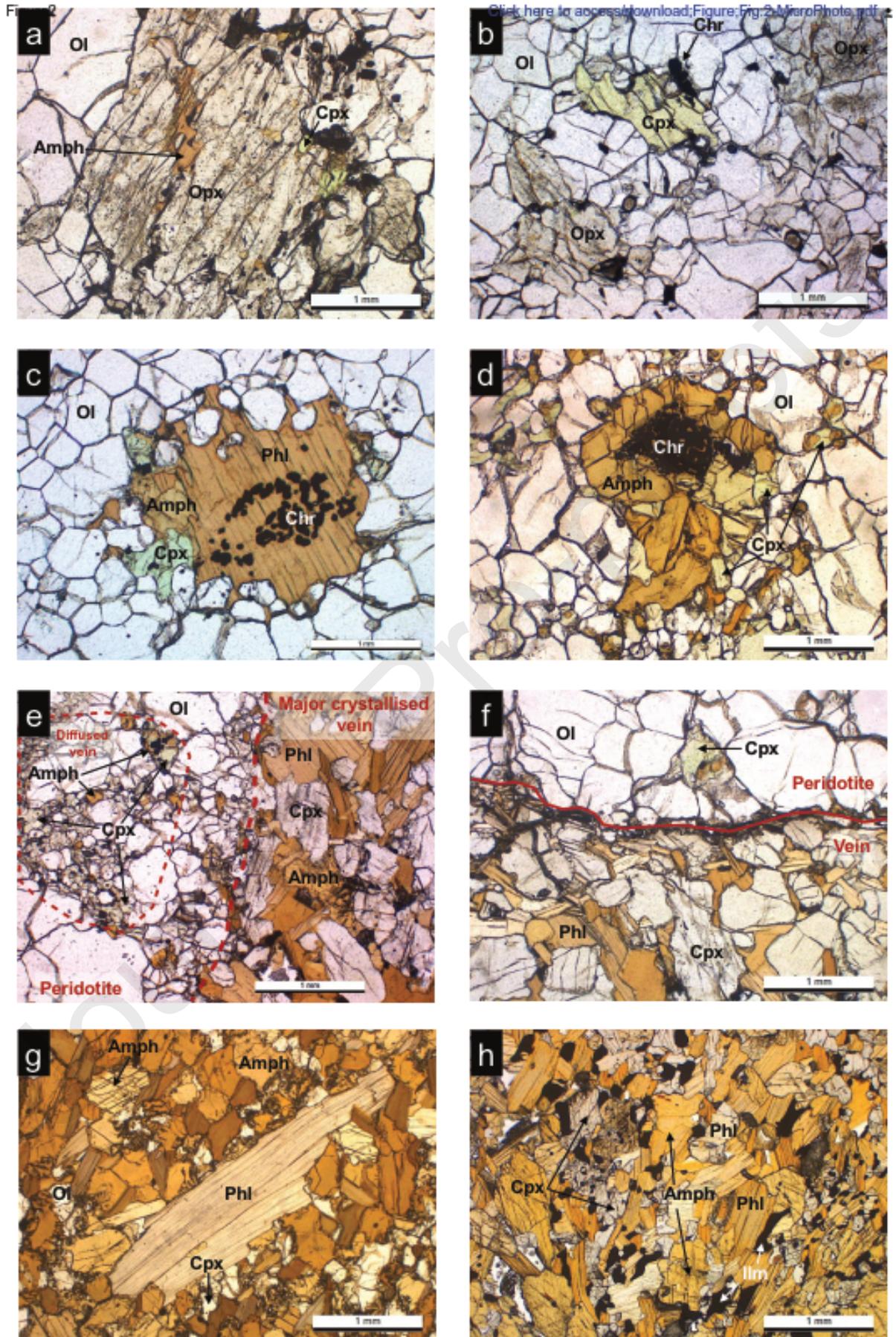
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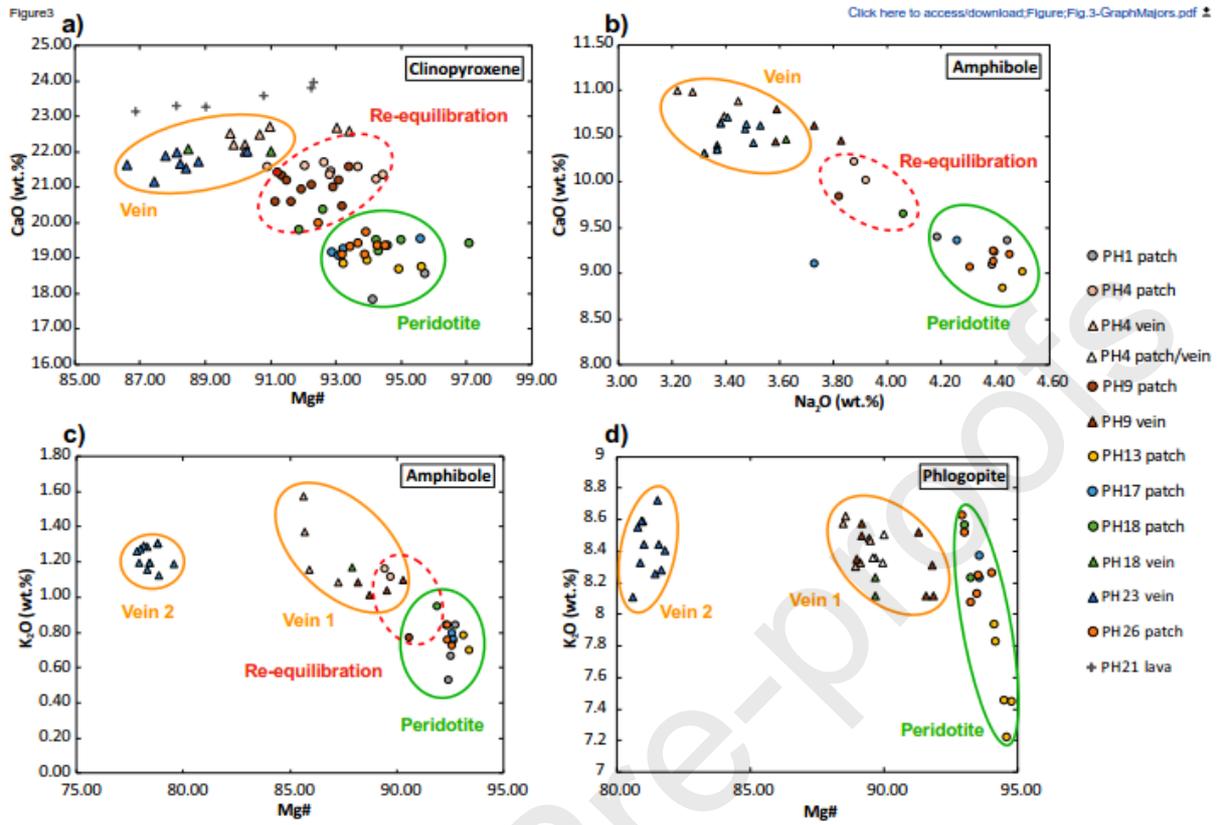
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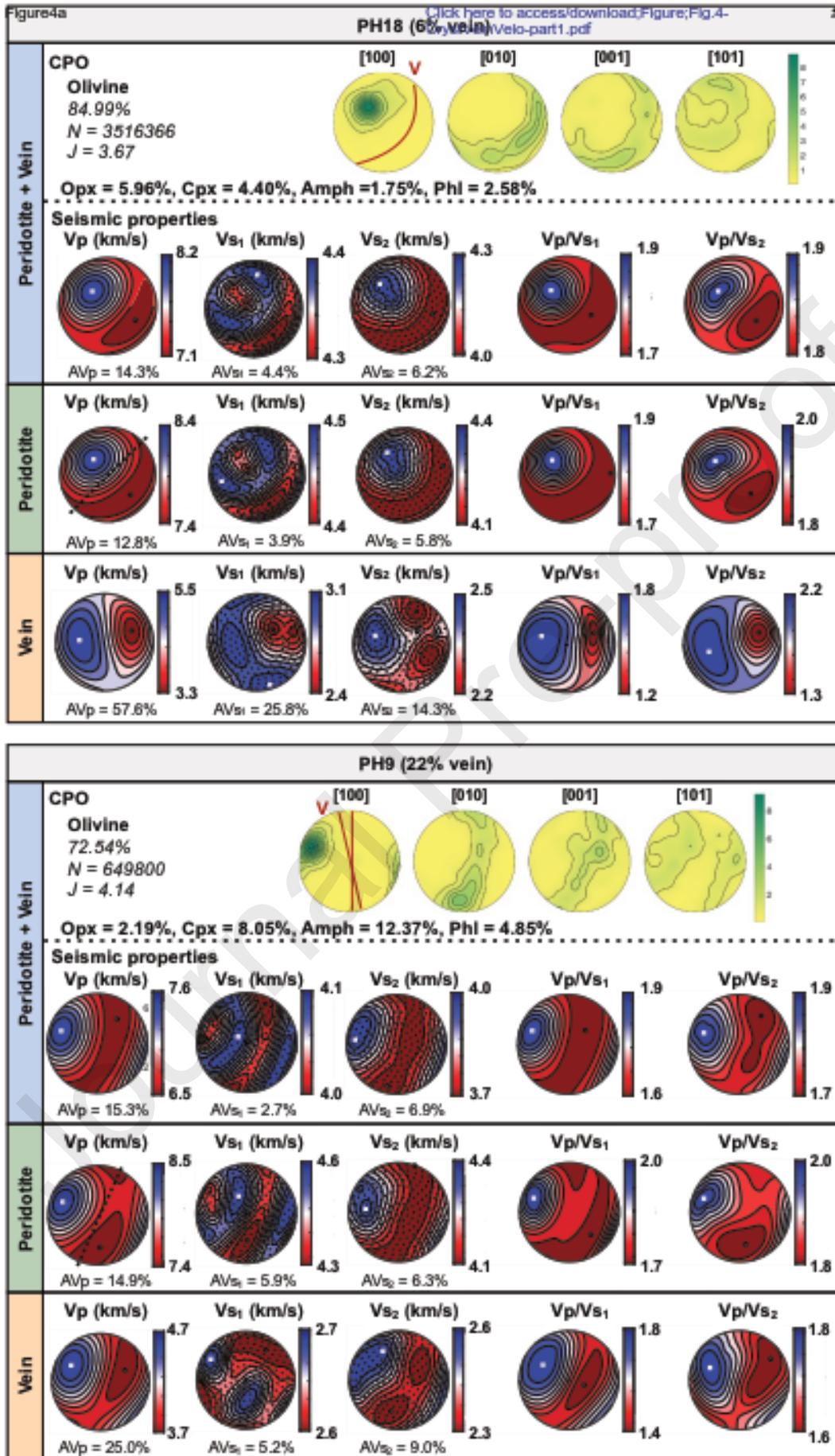
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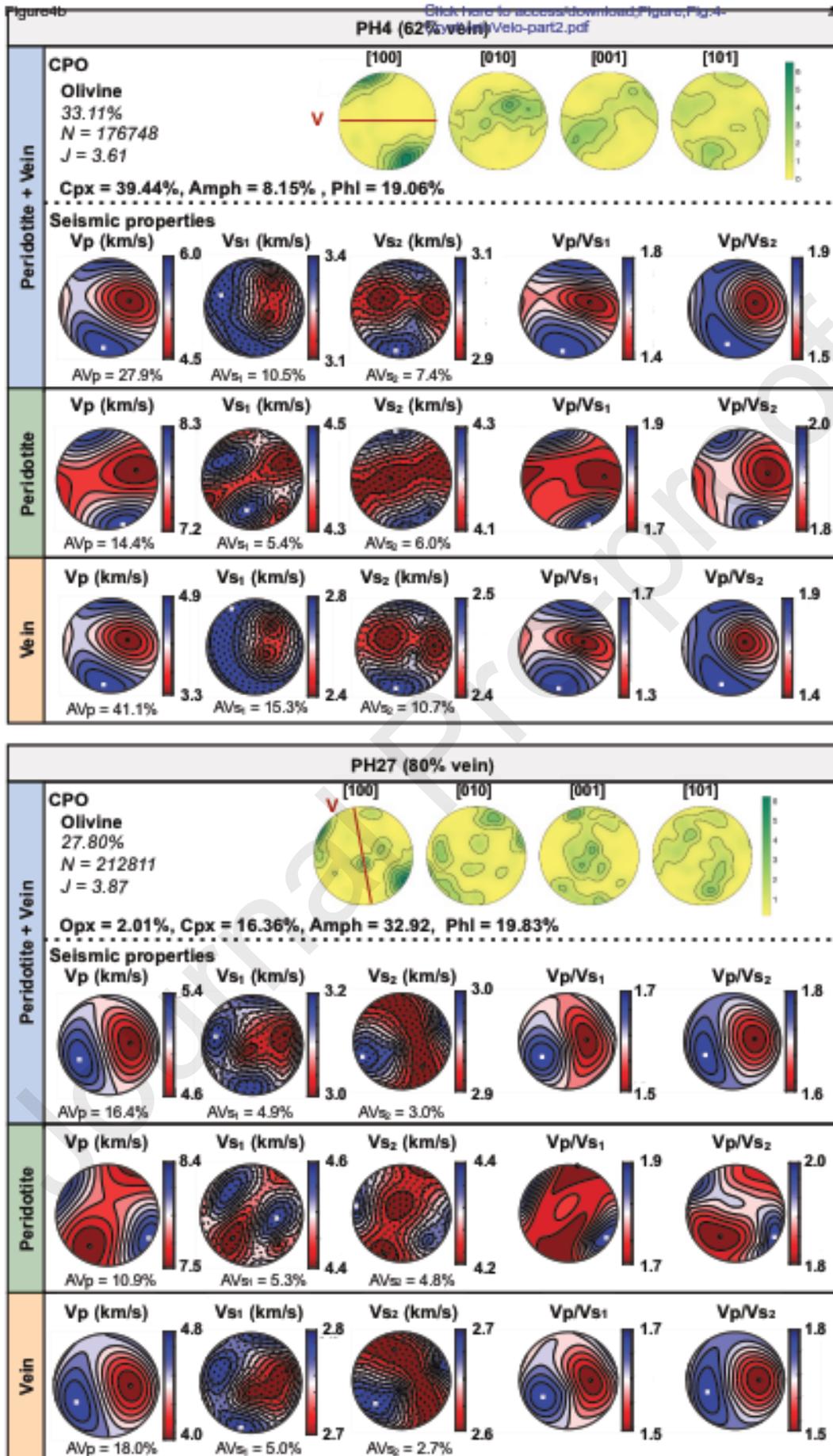
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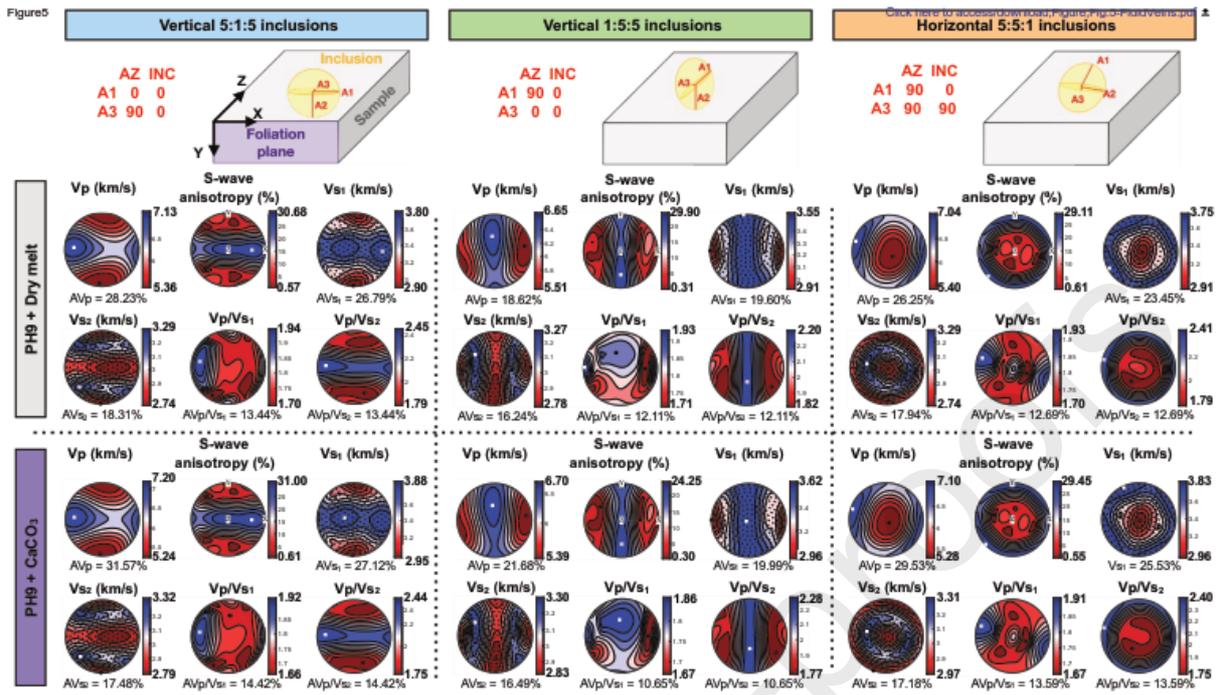
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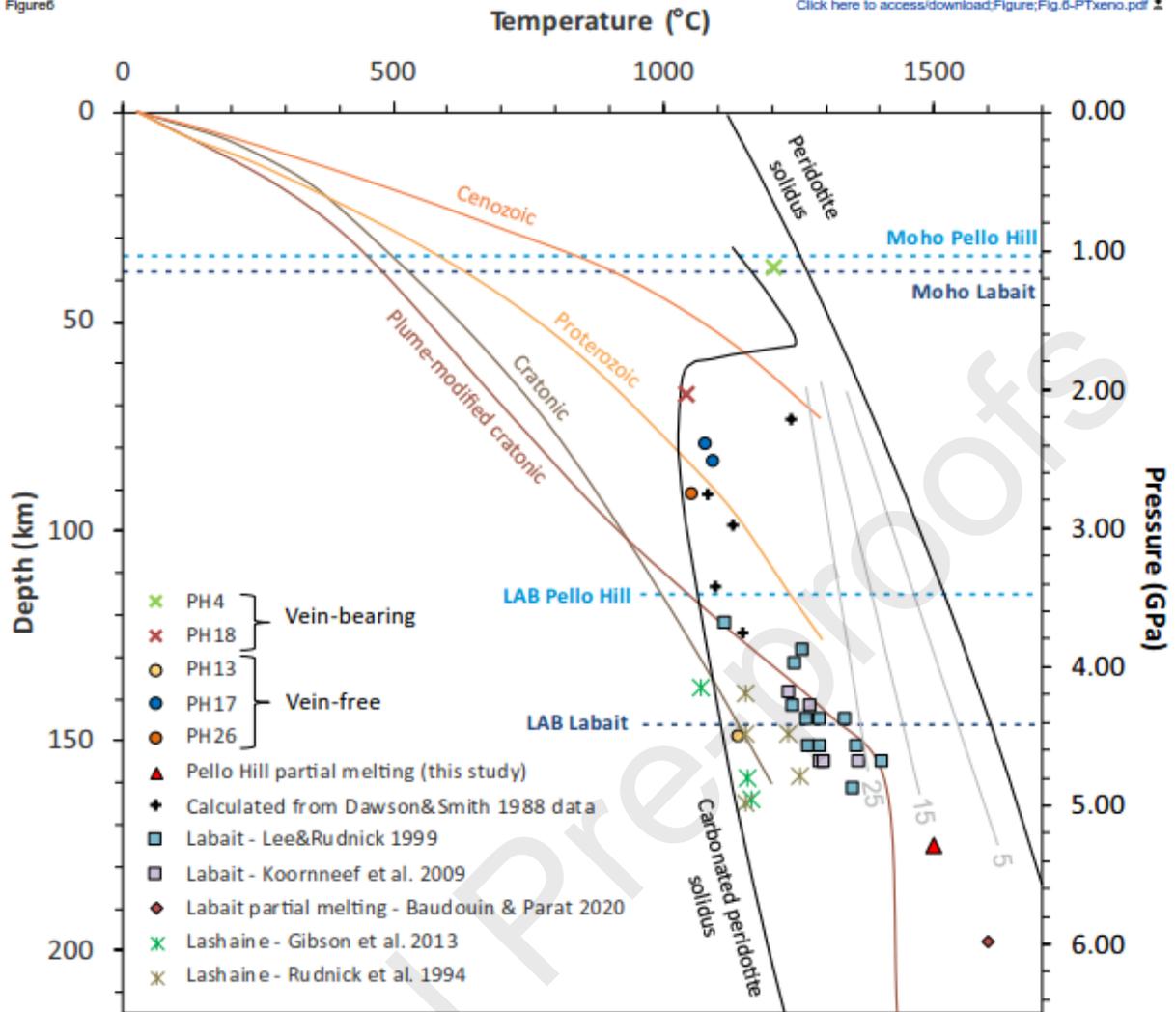
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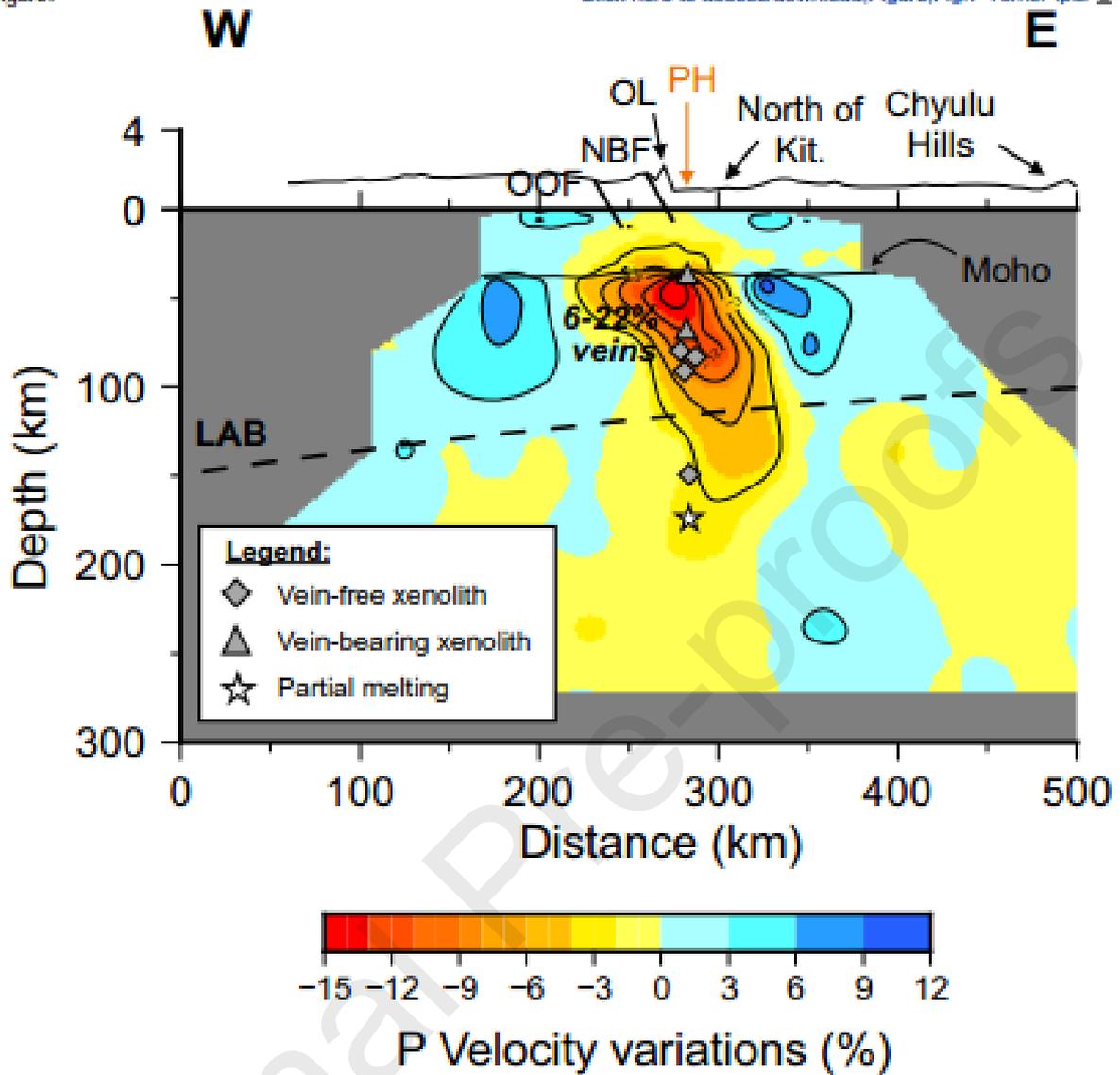
Figure 6

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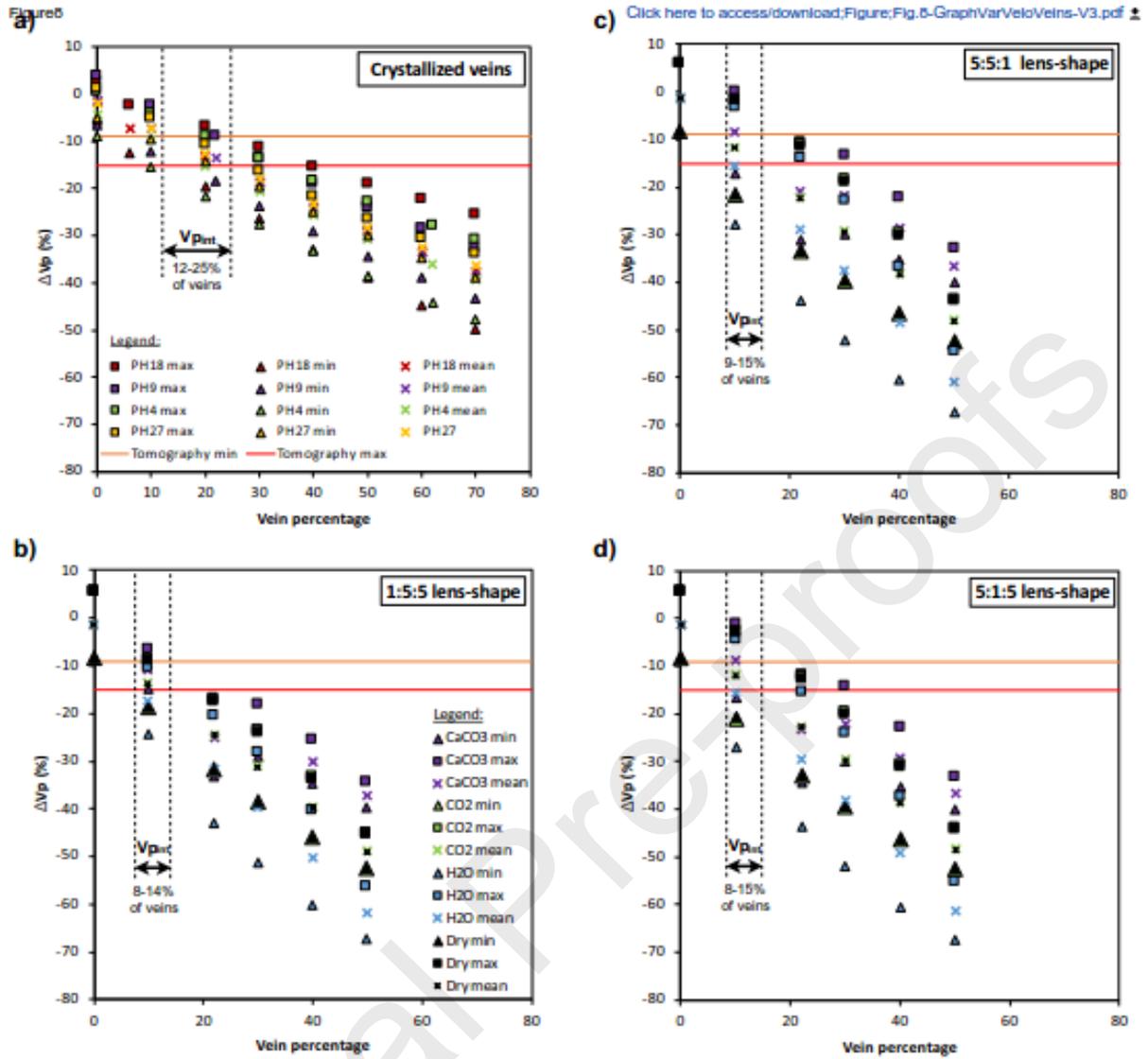
Figure 7

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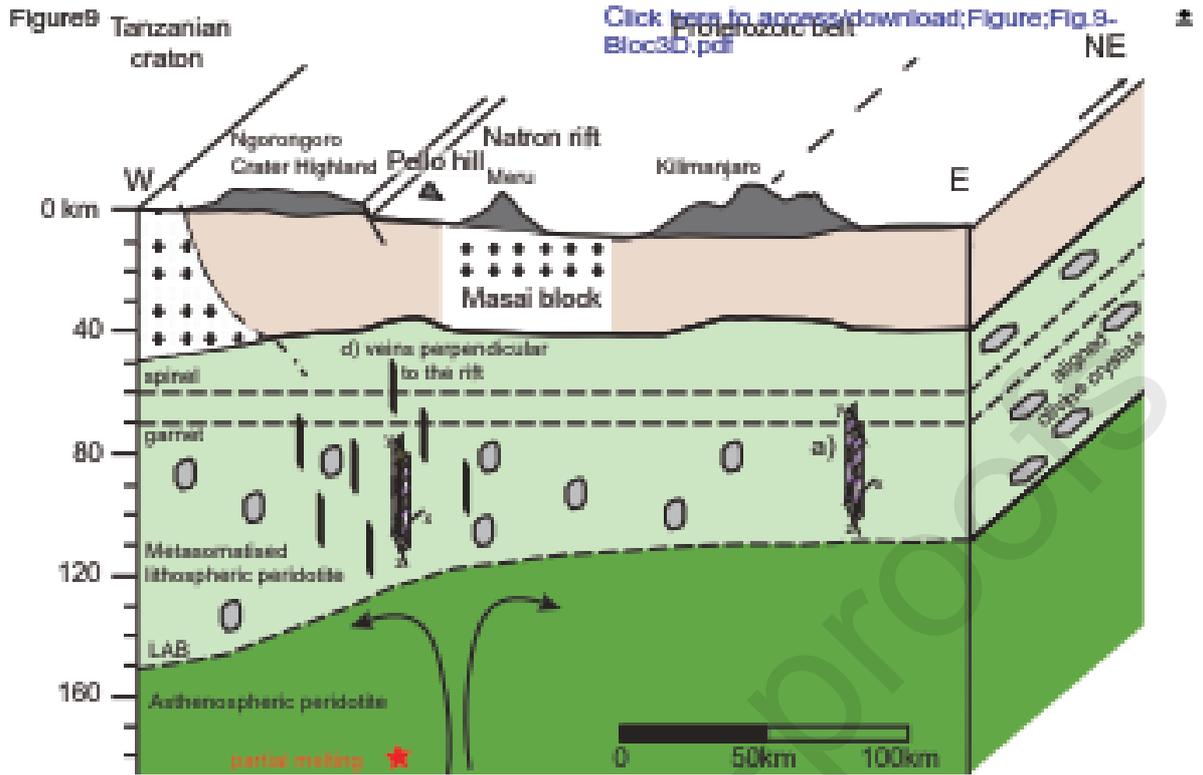
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Figure 8

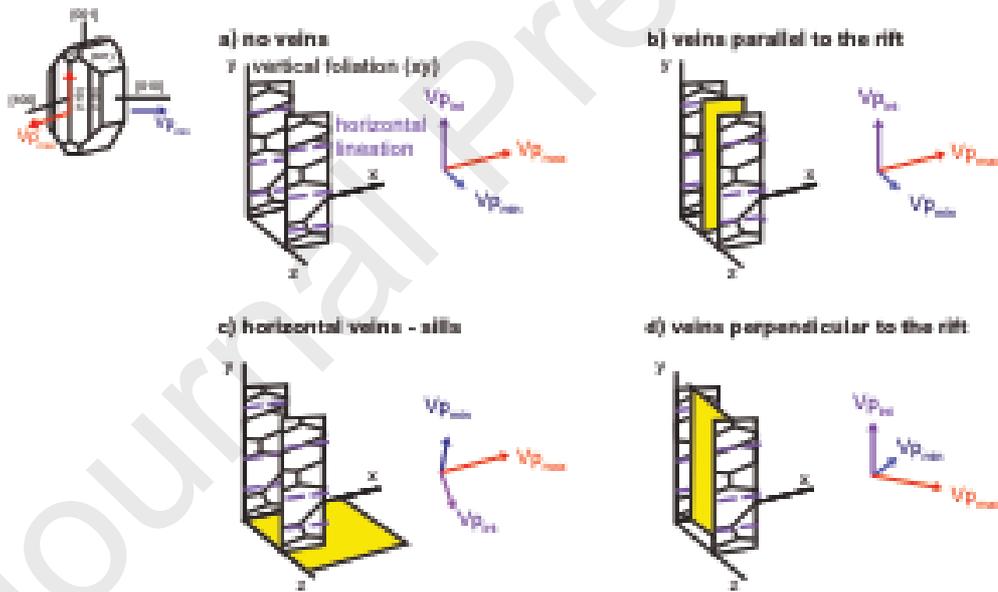


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Vertical foliation (xy) and horizontal lineation in peridotite



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