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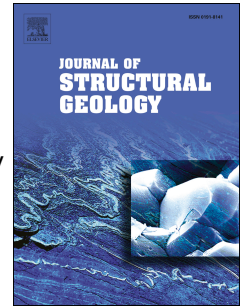
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1 **The internal deformation of the Peridotite Nappe of New Caledonia: a structural study**
2 **of serpentine-bearing faults and shear zones in the Koniambo Massif**

3
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13
14 **Keywords :** New Caledonia, ophiolite, serpentine, magnesite, nappe, structural analysis

15
16 **Abstract**

17 We present a structural analysis of serpentine-bearing faults and shear zones in the Koniambo
18 Massif, one of the klippen of the Peridotite Nappe of New Caledonia. Three structural levels
19 are recognized. The upper level is characterized by a dense network of fractures. Antigorite
20 and polygonal serpentine form slickenfibers along fault planes with distinct kinematics. As a
21 result, the upper level keeps the record of at least two deformation events, the first associated
22 with the growth of antigorite (WNW-ESE extension), the second with the growth of
23 polygonal serpentine (NW-SE compression). The lower level coincides with the ‘serpentine
24 sole’ of the nappe, which consists of massive tectonic breccias overlying a layer of mylonitic
25 serpentinites. The sole records pervasive tangential shear with top-to-SW kinematics and
26 represents a décollement at the base of the nappe. The intermediate level is characterized by

27 the presence of several meters-thick conjugate shear zones accommodating NE-SW
28 shortening. Like the sole, these shear zones involve polygonal serpentine and magnesite as the
29 main syn-kinematic mineral phases. The shear zones likely root into the basal décollement,
30 either along its roof or, occasionally, around its base. Compared to top-to-SW shearing along
31 the sole, the two deformation events recorded in the upper level are older.

32 The three structural levels correlate well with previously recognized spatial variations
33 in the degree of serpentinization. It is therefore tempting to consider that the intensity of
34 serpentinization played a major role in the way deformation has been distributed across the
35 Peridotite Nappe. However, even the least altered peridotites, in the upper level, contain so
36 much serpentine that, according to theoretical and experimental work, they should be nearly
37 as weak as pure serpentinite. Hence, no strong vertical gradient in strength due to variations in
38 the degree of serpentinization is expected within the exposed part of the nappe. Our proposal
39 is that strain localization along the serpentine sole results from the juxtaposition of the nappe,
40 made of weak serpentinized peridotites, against the strong mafic rocks of its substratum. This
41 interpretation is at odds with the intuitive view that would consider the nappe, made of
42 peridotites, as stronger than its basement.

43 **1. Introduction**

44 Ophiolites commonly occur within orogenic belts as a result of the obduction of a piece of
45 oceanic lithosphere upon a continental basement (Coleman, 1971; Dewey and Bird, 1971).
46 The ultramafic rocks that constitute a large part of most ophiolitic nappes generally display
47 evidence for pronounced fluid-rock interactions in the form of a spatially uneven development
48 of serpentinization. Due to their low strength in comparison to peridotites (e.g., Escartín et al.,
49 2001), serpentinites have the capacity to promote strain localization and greatly influence the
50 style of deformation on the lithosphere scale (e.g., Hirth and Guillot, 2013). Several types of
51 serpentine may be identified in a single ophiolite, which raises the question of their relative

52 chronology and their geodynamic environment of formation (e.g., Coulton et al., 1995). The
53 topic of this work is on ophiolitic nappes remaining near the surface, thus excluding the case
54 of dismembered meta-ophiolitic sequences occasionally exposed in the internal zones of
55 orogenic belts. In principle, serpentinization within an ophiolitic nappe may witness the
56 successive stages of oceanic accretion, intra-oceanic subduction (the ophiolite lying in the
57 upper plate, for instance in the fore-arc region), obduction, and post-obduction evolution.
58 Post-obduction events may involve post-convergence extension and/or collisional orogeny.

59 The Peridotite Nappe of New Caledonia is one of the few ophiolites in the world to
60 have escaped collision after obduction (e.g., Cluzel et al., 2012). In this case, deciphering
61 early serpentinization events should be easier. Recent work has shown that the Peridotite
62 Nappe hosts various types of serpentine (Lahondère and Maurizot, 2009; Ulrich, 2010;
63 Lahondère et al., 2012). An intimate association between deformation and the occurrence of
64 serpentines is recognized in these studies, however no attempt has been made so far to
65 characterize this deformation. More generally, structural studies carried out on the ultramafic
66 section of ophiolitic nappes usually focus on serpentine-free high grade fabrics (e.g., Boudier
67 et al., 1988; Suhr and Cawood, 1993; Cook et al., 2000). Studies including an analysis of
68 serpentine-bearing faults and shear zones are relatively rare (Bailey et al., 2000; Titus et al.,
69 2002; Schemmann et al., 2007; Laó-Dávila and Anderson, 2009; Federico et al., 2014).

70 In New Caledonia, earlier studies concerned with the Peridotite Nappe have focused
71 on i) the high temperature intra-oceanic stages of deformation within the nappe (Prinzhofer et
72 al., 1980; Titus et al., 2011), ii) the obduction process, however based essentially on the
73 analysis of the rock units underlying the nappe (e.g., Cluzel et al., 2001, 2012; Spandler et al.,
74 2005; Lagabrielle et al., 2013) and iii) post-obduction brittle extension that shaped the island
75 as a large horst (e.g., Lagabrielle et al., 2005; Chardon and Chevillotte, 2006). Furthermore,
76 much attention has been put on the chemistry and mineralogy of thick lateritic profiles formed
77 at the expense of the peridotites, which led to the development of worldclass nickel

78 mineralizations (e.g., Orloff, 1968; Trescases, 1975; Butt and Cluzel, 2013). In contrast, only
79 limited work has concerned serpentines-bearing faults and shear zones that are ubiquitous
80 throughout the Peridotite Nappe, despite the common statement that the distribution of nickel
81 ore is highly dependent on this fault network (Legu  r  , 1976; Cluzel and Vigier, 2008).

82 In this study, we had access to a recently opened large mining site in the Koniambo
83 Massif, one of the klippe of the Peridotite Nappe in northwestern New Caledonia (Fig. 1).
84 This site provides a quasi-continuous exposure of fresh rocks from the base of the nappe, near
85 sea level, up to ~800 m elevation, where the main laterite capping the peridotites occurs. We
86 focused on the study of structures and deformation associated with serpentines. These data
87 enable to describe the internal deformation of the nappe in relation with successive
88 serpentinization events. The implications of this study for the tectonic evolution of New
89 Caledonia are not discussed here and will be the topic of a distinct paper.

90 **2. Geological setting**

91 New Caledonia is located in the southwest Pacific Ocean, 1300 km east of Australia (Fig. 1).
92 On the main island, known as the “Grande Terre”, the Peridotite Nappe overlies, with a sub-
93 horizontal tectonic contact (Avias, 1967; Guillon, 1975), a substratum composed of several
94 volcano-sedimentary units (Paris, 1981; Cluzel et al., 2001, 2012). The Peridotite Nappe is
95 essentially exposed in the “Massif du Sud”, covering much of the southeastern third of the
96 Grande Terre, and as a series of klippe along the northwestern coast (Fig. 1). In the Massif
97 du Sud, the thickness of the nappe is at least 1.5 km and may reach 3.5 km (Guillon, 1975).
98 The nappe is mostly composed of harzburgites except in the northernmost klippe where
99 lherzolites dominate (e.g., Ulrich et al., 2010). Compositional layering is essentially
100 represented by 1 to 100 m-thick layers of dunite within the harzburgites. In the main part of
101 the nappe, the degree of serpentinization of the peridotites is variable but moderate (Orloff,
102 1968), serpentines occurring preferentially along a network of mm- to ~10 cm-thick fractures

103 and shear zones (e.g., Leguéré, 1976; Lahondère and Maurizot, 2009; Lahondère et al., 2012).
104 Fracture infillings and veins with various textures are described, containing
105 lizardite±chrysotile or antigorite±chrysotile and/or an additional type of light-coloured
106 serpentine not firmly identified (Lahondère and Maurizot, 2009; Lahondère et al., 2012).
107 Serpentinization becomes pervasive along the base of the nappe, forming a ‘sole’ in which
108 deformation has been intense (Avias, 1967; Orloff, 1968; Guillon, 1975; Leguéré, 1976;
109 Cluzel et al., 2012; Quesnel et al., 2013). The thickness of this sole is a few tens of meters in
110 the Massif du Sud but reaches a few hundred meters in the northwestern klippe (Guillon,
111 1975; Maurizot et al., 2002). According to Ulrich (2010), serpentinization of the sole involved
112 three stages, firstly a pervasive development of lizardite, secondly the formation of antigorite
113 within mm to cm-thick veins, finally the formation of chrysotile in local veinlet networks. In
114 some of the veinlets, Ulrich (2010) also reports the partial replacement of chrysotile by
115 polygonal serpentine. Recently, Ulrich et al. (2014) established that polygonal serpentine is
116 more widespread in the sole than previously thought, forming light green serpentine veins
117 associated with chrysotile veinlets.

118 Following their exposure to aerial conditions, the peridotites have been subject to
119 intense weathering under dominantly warm and wet climatic conditions. This has led to the
120 development of laterites up to ~30 m thick (e.g., Chevillotte et al., 2006). Leaching of the
121 peridotites by downward infiltrating meteoric waters has resulted in a redistribution of
122 elements, leading to high nickel concentrations at the base of the weathering profile, in the
123 transition zone between coarse-grained saprolite and the bedrock (Orloff, 1968; Trescases,
124 1975; Leguéré, 1976; Paris, 1981). Some veins with nickel-rich mineralizations show
125 evidence for a syn-tectonic emplacement (Cluzel and Vigier, 2008). Magnesite veins, which
126 frequently occur along the serpentine sole of the Peridotite Nappe, represent another by-
127 product of the leaching of the peridotites by meteoric fluids (Ulrich, 2010; Quesnel et al.,
128 2013). Quesnel et al. (2013) showed that at least some of these veins have been emplaced

129 during pervasive shearing along the sole.

130 In the Koniambo Massif, the Peridotite Nappe is exposed from its base, near sea level,
131 up to ~800 m elevation (Fig. 2a,b). Map relations (Carroué, 1972; Maurizot et al., 2002) show
132 that the basal contact is subhorizontal around much of the massif but delineates a large open
133 antiform in the west (Fig. 2b). At elevations above ~400 m, the nappe is capped by a highly
134 dissected and partly reworked lateritic profile (Maurizot et al., 2002). At lower elevations,
135 laterites of the westerly-inclined Kaféaté plateau probably belong to a younger planation
136 surface (Latham, 1977; Chevillotte et al., 2006). The Koniambo Massif essentially consists of
137 harzburgites with interlayers of dunite that define a crude compositional layering with a fairly
138 regular ENE-WSW strike and a ~50° southward dip (Maurizot et al., 2002). With regard to the
139 degree of serpentinization, Maurizot et al. (2002) distinguished three main rock types, namely
140 (i) moderately serpentinized peridotites in which the primary compositional layering is
141 preserved, (ii) highly serpentinized peridotites and (iii) massive serpentinites. The latter form
142 the serpentine sole of the nappe, about 200 m thick. The highly serpentinized peridotites
143 overlie the sole and form a distinct layer which, according to Maurizot et al. (2002), is ~200
144 m thick on the western flank of the massif but thins out further east (Fig. 2a,b). The
145 moderately serpentinized peridotites occupy the higher part of the massif.

146 3. Methods

147 3.1. Strategy

148 Field work focused on fresh rock exposures provided by the last ~5 years of mining activity in
149 the Koniambo Massif. The serpentine sole is well exposed in the Vavouto peninsula,
150 especially along two cross-sections here named VA1 (700 m long, striking N082°, Fig. 2c)
151 and VA2 (a composite section ~330 m long with a mean N-S strike). The VA2 section shows
152 the basal contact of the nappe and its substratum made of basalts and minor cherts (Fig. 3a).
153 Further east, the mine access road runs from the base of the nappe (height spot 65 m in Figure

154 2a) to the upper part of the massif; it provides good outcrops of the intermediate level of
155 highly serpentinized peridotites. The moderately serpentinized peridotites are exposed at
156 higher levels of the access road and in a series of currently exploited open pits. We carried out
157 preliminary field observations and sampled rocks with various textures and fabrics (veins,
158 fracture infillings, fault coatings, cleavage domains within shear zones). We determined the
159 mineralogical content of the samples, which then served as references for macroscopic
160 assessment of the mineralogy within specific structures during renewed field work and
161 structural analysis.

162 3.2. Analytical methods

163 Reference samples have been characterized through a multi-technique approach including
164 transmission electron microscopy, microprobe analysis (not presented here) and Raman
165 spectroscopy. Raman spectroscopy measurements were carried out at the laboratory
166 GéoRessources Nancy, France, using a Horiba Jobin-Yvon Labram HR800 spectrometer and
167 a visible ionized argon laser source with a wavelength of 514 nm. The output power of the
168 laser was 100 mW and measurements were performed using an Olympus lens of x50 to focus
169 the laser beam onto an area 1 μ m in diameter. Spectra are the average of 6 to 10 acquisitions
170 of 20 s each to optimize the signal/noise ratio. For serpentine identification, only one region
171 of the Raman spectrum has been investigated, between 3520 cm^{-1} and 3870 cm^{-1} (Fig. 4). This
172 region enables to characterize the hydroxyl groups which are the most discriminant for
173 distinguishing the polymorphs of serpentine (Auzende et al., 2004; Schwartz et al., 2013;
174 Ulrich et al., 2014).

175 For analyzing the deformation, we focused our attention on shear zones and fault
176 planes. The kinematics of shear zones have been determined using the obliquity of cleavage
177 with respect to the shear zone walls (Fig. 5e,f) and, occasionally, the occurrence of a set of
178 oblique shear bands. For determining the kinematics of faults, the offset of pre-kinematic

179 markers or the presence of subsidiary Riedel shears has occasionally been used, however in
180 most cases we favoured the use of accreted slickenfibers with a clear staircase geometry (Fig.
181 5a,b,c,d). The advantage of stepped slickenfibers is to provide a robust sense-of-slip indicator
182 and to allow establishing which mineral or mineral assemblage was stable during a given fault
183 slip event (Durney and Ramsay, 1973; see also, e.g., Twiss and Moores, 1992).

184 The bulk deformation of densely faulted areas has been assessed using the FaultKin
185 software of Allmendinger et al. (2012). The basic tenets of this software, described by Marrett
186 and Allmendinger (1990), are the same as those of the right-dihedra method (Angelier and
187 Mechler, 1977; Angelier, 1994). It should be noticed that the approach used by Marrett and
188 Allmendinger (1990) is intended to provide a qualitative description of the strain ellipsoid, not
189 the stress ellipsoid. Several authors have argued that, in essence, the analysis of fault slip data
190 informs more about strain than about stress (e.g., Twiss and Unruh, 1998; Tikoff and Wojtal,
191 1999; Gapais et al., 2000). According to C  lerier et al. (2012), methods aiming at retrieving
192 paleostresses may be more suited for separating distinct subsets (presumed to witness
193 successive tectonic events) within a given fault slip dataset whereas methods aiming at
194 retrieving strain (e.g., Marrett and Allmendinger, 1990) may be more suited for yielding an
195 average estimate of strain within a rock volume. In our case, the mineralogy of slickenfibers,
196 which is of two types, has provided an independent criterion for separating the fault slip data
197 into two subsets, therefore the ‘strain’ approach seems more appropriate.

198 **4. Results**

199 4.1. Reference samples

200 We here describe three hand specimens that are representative of field cases where serpentine
201 occurs in association with deformation. The two first specimens come from higher levels of
202 the Koniambo Massif. The third sample comes from the Vavouto peninsula, within the
203 serpentine sole.

204 The first sample includes a ~1 cm-thick vein of light olive green serpentine with a
205 massive aspect (Fig. 4a). The vein has been active as a fault plane as indicated by fine
206 striations underlined by black oxides along a surface parallel to its margin. The Raman
207 spectrum of the vein material (analysis #1-1) shows two closely spaced main bands centered
208 on $\sim 3688\text{ cm}^{-1}$ and $\sim 3694\text{ cm}^{-1}$, which are diagnostic of polygonal serpentine (Auzende et al.,
209 2004; Ulrich, 2010; Ulrich et al., 2014).

210 The second sample includes a ~1 cm-thick vein of platy-fibrous serpentine (Fig. 4b).
211 Outside the vein, the host rock shows a Raman spectrum (#2-1) with two main bands centered
212 on $\sim 3688\text{ cm}^{-1}$ and $\sim 3696\text{ cm}^{-1}$, indicating the presence of polygonal serpentine. The vein
213 itself is composite. Two insulated domains with diffuse boundaries are made of darker
214 serpentine showing a Raman spectrum (#2-2) with two main bands centered on $\sim 3672\text{ cm}^{-1}$
215 and $\sim 3700\text{ cm}^{-1}$, typical for antigorite. The main part of the vein appears as an heterogeneous
216 stack of fibers. The Raman spectrum of analysis #2-3 in this domain shows a minor band
217 centered on $\sim 3671\text{ cm}^{-1}$, attesting for the presence of antigorite. In addition, the main band
218 centered on $\sim 3698\text{ cm}^{-1}$ and the adjacent poorly developed band at $\sim 3688\text{ cm}^{-1}$ indicate the
219 presence of polygonal serpentine or/and chrysotile. Finally, the margin of the vein is
220 underlined by small domains of homogeneous light green serpentine. Their staircase
221 distribution indicates that they formed as the result of a fault slip event. The Raman spectrum
222 of analysis #2-4 shows a main band centered on $\sim 3697\text{ cm}^{-1}$ and an adjacent poorly developed
223 band at $\sim 3690\text{ cm}^{-1}$ which, again, points to the presence of polygonal serpentine or/and
224 chrysotile. In that case, however, chrysotile is unlikely because the serpentine in this part of
225 the vein is massive and does not display the habitus of well defined submillimetric fibers that
226 is typical for chrysotile (Lahondère and Maurizot, 2009; Ulrich, 2010; Lahondère et al.,
227 2012). The small domains of polygonal serpentine have well defined boundaries, which
228 suggests that they formed later than the adjacent domain of darker antigorite did.
229 Furthermore, the platy-fibrous nature of the main part of the vein is typical for antigorite (e.g.,

230 Lahondère et al., 2012), therefore the occasional intimate association of antigorite and
231 polygonal serpentine, like in analysis #2-3, can be interpreted as the result of a partial
232 pseudomorphosis of antigorite by fine-grained polygonal serpentine. Thus, the crystallization
233 of polygonal serpentine appears to postdate that of antigorite.

234 The third sample consists of a fist-sized elongate clast taken from one of the major
235 shear zones met along the VA1 section (Fig. 2c). The clast, which is made of pervasively
236 serpentinized harzburgite with few orthopyroxene relics, is embedded in a film up to 0.5 cm
237 thick of homogeneous pale green serpentine (Fig. 4c,d). The Raman spectrum (analysis #3-1)
238 shows two closely spaced main bands centered on $\sim 3691\text{ cm}^{-1}$ and $\sim 3697\text{ cm}^{-1}$, indicating that
239 the film consists of polygonal serpentine.

240 4.2. Field determination of the serpentine polymorphs associated with deformation

241 Based on the characterization of the three above samples as well as other reference samples,
242 we have established a number of macroscopic criteria for a field assessment of the mineralogy
243 of fault zones and shear zones. Within fault zones, antigorite and polygonal serpentine are
244 both able to build macroscopic slickenfibers with a clear staircase geometry. For
245 distinguishing the two polymorphs, the main argument is textural. Antigorite has a well
246 expressed platy-fibrous aspect (see also, e.g., Lahondère et al., 2012) with fibers frequently
247 several centimeters long (Fig. 5a). In contrast, polygonal serpentine is massive, matte, and
248 develops only short slickenfibers (Fig. 5b,c). A second argument is the fact that polygonal
249 serpentine is pale green whereas antigorite is generally darker (Figs. 4b,c,d and 5a,b,c).
250 However, in the saprolitic levels, weathering modifies the colours and polygonal serpentine
251 tends to become olive green (Figs. 4a and 5c). In addition, some fault planes bear
252 slickenfibers showing mixed properties, with a clear platy-fibrous nature but a pale green
253 colour and a matte aspect (Fig. 5d). In that case, taking into account the likeliness of a
254 pseudomorphic replacement of antigorite by polygonal serpentine, as in the second reference

255 sample above, we have assumed that the slickenfibers witness a syn-antigorite slip event.

256 Antigorite and polygonal serpentine are also involved in shear zones. There, antigorite
257 forms distinct lamellae (Fig. 5e) whereas polygonal serpentine is more massive despite the
258 development of a pronounced cleavage (Fig. 5f). Most of the shear zones we have observed
259 involve polygonal serpentine. There is always a close spatial coincidence between the shear
260 zone walls and the margins of the serpentine body (Fig. 5e,f). The fabric in each shear zone is
261 pronounced and, in principle, shearing might post-date serpentinization. Nevertheless, would
262 shearing occur under physical conditions outside the stability range of a primary serpentine, it
263 is expected that a new serpentine polymorph would easily crystallize during the pervasive
264 deformation. Hence, for shear zones with an homogeneous serpentine content, we assume that
265 shearing and mineral growth were coeval.

266 Lizardite and chrysotile were also observed in the field, with features similar to those
267 reported by Ulrich (2010) and Lahondère et al. (2012). Dark lizardite is widespread, present
268 as a diffuse grain-scale network in the peridotites and as infillings of ~1 mm to ~5 cm-wide
269 joints. It is also found on the margins of many of the faults and shear zones that involve
270 antigorite or polygonal serpentine (Fig. 5c,e). However, lizardite itself only rarely bears
271 striations and, in our experience, never builds macroscopic slickenfibers, therefore we made
272 no attempt to characterize the deformation associated with lizardite. Chrysotile occurs as very
273 thin fibers always oriented at right angle to the vein walls within ≤ 1 cm-thick veinlets. Dense
274 networks of subparallel veinlets are locally observed. Such veins can be interpreted as tension
275 gashes, therefore, in principle, they could provide some clue on the tectonic regime that
276 prevailed during chrysotile growth. Alternatively, tension gashes may open in the absence of
277 a significant tectonic contribution, for instance due to fracturing assisted by fluid pressure.
278 For this reason, and because these thin veins represent very small amounts of strain, we
279 ignored them during the analysis of deformation.

280 4.3. Structural analysis

281 We identified three structural levels, which correlate with the rock horizons identified by
282 Maurizot et al. (2002) on the basis of the intensity of serpentinization (Fig. 2a,b).

283 4.3.1 Upper structural level

284 The Upper structural level coincides with the domain of moderately serpentinized
285 peridotites at higher levels of the massif. A dense network of fractures, most of them dipping
286 steeply, crosscut the peridotites (Fig. 6a). The spacing between adjacent macroscopic
287 fractures is always less than ~2 m. In between fractures, the least serpentinized peridotites
288 show 'mesh' textures (e.g., Rouméjon and Cannat, 2014) resulting from a partial replacement
289 of olivine by lizardite along grain boundaries and grain-scale microfractures (Fig. 6b). On the
290 outcrop scale, in addition to 'dry' joints, many fractures are filled with one or several types of
291 serpentine. When present together with another serpentine polymorph, lizardite occupies the
292 margins of the fracture (Fig. 5c). When occurring together, antigorite and polygonal
293 serpentine occupy the core vs. the margins of the fracture, respectively. As discussed with the
294 second reference sample in Section 4.1 (Fig. 4b), polygonal serpentine formed later.

295 In most veins involving fibrous polygonal serpentine or/and antigorite, the fibers lie at
296 a very low angle to the vein walls, hence these veins can be interpreted as fault zones (e.g.,
297 Durney and Ramsay, 1973). A total of 125 such faults has been measured in a series of open
298 pits covering a relatively small area on the top of the Koniambo Massif (Fig. 2a). The syn-
299 polygonal serpentine vs. syn-antigorite nature of each fault displacement has been determined
300 by assessing the mineralogy of associated slickenfibers (Fig. 5a,b,c,d) (see Section 4.2). As a
301 result, two fault subsets have been defined, which compare well in terms of orientation but
302 have distinct kinematics (Fig. 7a,b). The majority of the faults with dominantly dip-slip
303 movement have a normal sense of slip when associated with antigorite vs. a reverse sense of
304 slip when associated with polygonal serpentine. Fault slip analysis documents two strikingly

305 different strain ellipsoids (Fig. 7c,d). The syn-antigorite faults yield a highly constrictional
306 ellipsoid, as illustrated by the value of 0.17 computed for the R ratio ($R=(\epsilon_2-\epsilon_3)/(\epsilon_1-\epsilon_3)$).
307 Consequently, λ_1 , the axis of maximum stretching, is the most accurately defined; it lies
308 horizontally with a N102° trend (Fig. 7c). The axis of maximum shortening, λ_3 , plunges
309 steeply (64°) to the north-northeast. The syn-polygonal serpentine faults yield an ellipsoid in
310 the flattening field ($R=0.71$). Consequently, λ_3 is the most accurately defined axis; it lies
311 subhorizontally with a N144° trend, while λ_1 plunges shallowly in the N057° direction (Fig.
312 7d). Figure 8 illustrates the results of fault analysis on the scale of a pit ~200 m wide within
313 the main area of investigation (Pit 208) and on an isolated site located 2 km further northwest
314 (Bilboquet, see Figure 2a for location). Fault slip analysis on these sites yields results that fit
315 closely with those deduced from the bulk dataset.

316 Further southwest and at lower elevation with respect to the open pits, the mine access
317 road provides a large outcrop near the transition between the upper and intermediate structural
318 levels (Roadsite '620 m' in Figure 2a). It shows fault planes with slickenfibers as well as
319 shear zones a few centimeters thick (Fig. 5f), both involving polygonal serpentine. Fault slip
320 analysis yields an ellipsoid in the flattening field ($R=0.66$), with λ_3 shallowly plunging
321 southward (27°) along a N006° trend (Fig. 9). This is broadly consistent with the syn-
322 polygonal serpentine strain field deduced from the open pits (Fig. 7d), though with a ~40°
323 clockwise rotation in the direction of λ_3 . This difference could reflect differential large-size
324 block rotations post-dating the measured fault slip increments. However, in this area, dunitic
325 interlayers retain the same ENE-WSW strike and ~50° southward dip as elsewhere in the
326 Koniambo Massif (Maurizot et al., 2002), therefore this hypothesis is unlikely. Hence, the
327 difference is more likely to reflect a modification of the strain field through space or time.

328 4.3.2 Intermediate structural level

329 In the Intermediate structural level, the degree of serpentinization is overall high but

330 laterally variable. The main part is similar to exposures of the Upper structural level, with a
331 dense network of serpentine-bearing fractures. The distinctive feature of the Intermediate
332 structural level is that this rock mass is crosscut by low-dipping shear zones a few meters
333 thick, characterized by a pervasive development of light green polygonal serpentine (Fig.
334 10a). Two such shear zones (SZ1 and SZ2) occur at high levels of the Intermediate structural
335 level (Fig. 2a). The NE-SW horizontal spacing between the two shear zones is ~450 m.
336 Continuous rock exposure along the mine access road enables to note that no similar shear
337 zone exists between SZ1 and SZ2. Both SZ1 and SZ2 involve polygonal serpentine with a
338 closely spaced cleavage as well as magnesite (Fig. 10b,c,d,e). In the case of SZ1, striations
339 have not been observed. The direction of shearing is assumed to lie at right angle to the
340 intersection between the cleavage and shear planes located along the margins of the shear
341 zone (Fig. 10b,c), along a ~N034° trend (Fig. 10f). In the case of SZ2, a major fault plane in
342 the core of the shear zone bears striations with a N039° trend while the relationship between
343 cleavage and shear bands suggests a ~N085° trend for the direction of shearing (Fig. 10e,f).
344 Both shear zones have top-to-SW kinematics. SZ1 has the attitude of a reverse-slip shear zone
345 (Fig. 10a,b,c,f) while SZ2 combines reverse and dextral displacements (Fig. 10e,f).

346 At lower elevations, the mine access road crosscuts a prominent SW-dipping fault
347 zone involving a ~30 m-thick pinch of fine-grained sediments (Maurizot et al., 2002) (Fig.
348 11a). These sediments are originally part of either the 'Poya terrane' (Maurizot et al., 2002)
349 or, more likely, the 'Nepoui flysch' (D. Cluzel, personal communication, 2012). As a result,
350 they represent a fragment of the volcano-sedimentary series that immediately underlie the
351 Peridotite Nappe (Cluzel et al., 2001). Hence, the fault zone likely roots slightly beneath the
352 basal contact of the nappe, or along it if one takes into account the fact that sheets of
353 serpentinite are frequently mixed with substratum rocks around several klippen of the nappe
354 (e.g., Maurizot et al., 1985). Within the fault zone, the main fault superposes highly sheared
355 serpentinites onto the sediments (Fig. 11a). Its orientation is N146°, 38°SW and it bears

356 pronounced striations with a N072° trend (Fig. 11b). Along this contact and across a structural
357 thickness of at least 75 m above it, shear sense criteria indicate a reverse-slip (top-to-ENE)
358 displacement (Fig. 11c,d). This movement is consistent with the incorporation of rocks from
359 the substratum into the fault zone, and with the probable offset of the roof of the serpentine
360 sole across it (Fig. 2a,b). This top-to-ENE fault can be viewed as a conjugate shear with
361 respect to the top-to-SW SZ1 and SZ2 shear zones.

362 4.3.3 Lower structural level (the serpentine sole)

363 The best exposures of the serpentine sole lie in the Vavouto peninsula (Figs. 2, 3, 12
364 and 13). Further east, the map contour of the basal contact of the nappe implies a culmination
365 of this contact slightly northwest of the Tambounan Peak (Fig. 2a). As a result, the surface
366 probably dips westward beneath the peninsula, with a slope of ~6-7° (Fig. 2b). The contact is
367 exposed at the eastern margin of the peninsula, displaying ~E-W-trending open folds on a
368 ~100 m scale (Fig. 3a).

369 The rocks along the sole are brecciated and/or foliated. Breccias are well developed
370 along the VA1 section (Fig. 12a) and consist of clasts of serpentinized peridotite separated by
371 serpentine joints (Fig. 4d). Many joints bear very fine striations (Fig. 4c). The striations have
372 extremely variable orientations, which probably results from the complexity of clast
373 displacements/rotations during brecciation. Major shear zones occur along the VA1 section
374 (Fig. 2c), marked by a reduction in clast size (Figs. 12a and 13a). The shear zone boundaries
375 are dominantly diffuse, but locally change to sharp. The width of the main shear zones ranges
376 from ~2 to ~10 m. In the shear zones, the greater density of fault planes subparallel to their
377 walls tends to produce elongate clasts that locally define a crude foliation (e.g., in the upper
378 part of Figure 13b). In the core of some shear zones, the planar fabric is more pronounced and
379 locally oblique to the shear zone walls. In cases where this oblique fabric had the appearance
380 of a closely spaced cleavage, it helped to assess the kinematics of the shear zone. In most

381 cases, however, it is difficult to decide whether the fabric could alternatively represent a dense
382 pattern of Riedel shears, hence implying the opposite sense of shear compared to that deduced
383 from an oblique cleavage. As a result, the kinematics of individual shear zones and shear
384 planes along the VA1 section has essentially been deduced from their effect (offset and drag
385 folding) on pre-existing markers (Figs. 12b and 13b).

386 The largest part of the VA2 section consists of serpentinites with a pronounced planar
387 fabric (Figs. 3a and 12c). On a 10 cm scale, this fabric is underlined by a disjunctive
388 anastomosing cleavage (Fig. 13c). When present, clasts are isolated, hence the rock retains a
389 matrix-supported texture (Fig. 12d). Generally, the clasts have an elongate and subangular to
390 rounded shape with a consistent orientation throughout the rock (Fig. 12d). As a result, the
391 rock has the typical appearance of a mylonite. Early compositional layering has been
392 transposed into parallelism with the cleavage (Fig. 12c), attesting for very large strains. Under
393 microscope, the fabric of the foliated serpentinites is irregularly distributed but locally
394 pervasive (e.g., in the right part of Figure 13d). A component of ductile deformation is thus
395 involved. The northern end of the VA2 section shows that the foliated serpentinites occur
396 immediately above the basal contact of the nappe (Fig. 3a). The southern end of the section
397 shows that they underlie the part of the sole dominated by breccias through a relatively sharp
398 transition zone (Fig. 3b). Hence, the foliated serpentinites form a distinct layer of most
399 intensely deformed rocks at the base of the serpentine sole, with a thickness of at least 20 m.

400 As illustrated by the third reference sample in Section 4.1 (Fig. 4c,d), our observations
401 indicate that the major shear zones along the VA1 section involve pale green polygonal
402 serpentine as the main syn-kinematic mineral phase (Figs. 12a and 13a,b). This also applies to
403 the layer of foliated serpentinites of the VA2 section (Figs. 12c,d and 13c). Magnesite also
404 occurs within this layer and along the VA1 section, in the form of irregularly distributed veins
405 and local stockworks (Figs. 2c, 12c and 13a). Magnesite occurs in close association with

406 polygonal serpentine (Figs. 12c,d and 13a,b,c). Evidence of the syn-kinematic nature of the
407 magnesite veins is shown in Figure 13 (see also Quesnel et al., 2013). In the macroscopic
408 examples b and c, the veins are offset by local shear planes that contribute to the bulk
409 shearing deformation. In addition to the vein infilling habitus, elongate clusters of partly
410 coalescent millimetric to centimetric nodules of magnesite are occasionally found. Figure 13d
411 shows part of a pluricentimetric nodule under microscope. Its right margin is irregular, made
412 up of submillimetric nodules of cryptocrystalline magnesite. This geometry, which suggests
413 an unconstrained growth of magnesite, could be taken as an indication that magnesite has
414 formed in the absence of contemporaneous deformation, after the development of the shear
415 fabric visible in the host serpentinite. However, the upper margin of the nodule is straight and
416 coincides with one of the shear bands related to this fabric. The relations between the nodule
417 and the shear band, highlighted in the enlargement, are clearly more in line with the shear
418 band post-dating the nodule than with the opposite. Overall, the relations seen in Figure 13d
419 are consistent with magnesite having grown during shearing.

420 The asymmetric distribution of shear zones along the VA1 (Fig. 2c) and VA2 (Fig.
421 12c) sections documents non coaxial deformation across the serpentine sole. The exposure
422 surface being smooth along both sections, the orientation of the main shear zones is difficult
423 to evaluate. Nevertheless, a dozen shear planes of significant size (several meters) have been
424 measured along the VA1 section (Fig. 12e). Their mean orientation is N141°, 46° SW. This
425 suggests top-to-SW kinematics for the bulk shearing deformation. Top-to-SW shearing is
426 consistent with the orientation of steep planar magnesite veins in between the shear zones
427 (mean value N134°, 62°SW), interpreted as tension gashes (Quesnel et al., 2013), and with the
428 orientation of folds developed at the expense of magnesite veins within the shear zones (Fig.
429 12e). Top-to-SW shearing is also consistent with the apparent top-to-west vs. top-to-south
430 sense of shear observed along the ~W-E-trending VA1 vs. ~N-S-trending VA2 section,
431 respectively. The orientation of the ellipsoid delineated by the large ovoid clast in the upper

432 left corner of Figure 3a also fits with this interpretation. Along the VA1 section, the major
433 shear zones with an apparent top-to-west sense of shear have a mean apparent dip of 24.5°
434 (Fig. 2c). Assuming that the mean strike of these shear zones is $N141^\circ$ (i.e. the same as for the
435 meter-scale shear planes) and taking into account the $N082^\circ$ strike of the VA1 section, the
436 mean true dip of the major shear zones can be estimated at $\sim 28^\circ$. Taking into account the
437 likely $\sim 6\text{-}7^\circ$ westward slope of the basal contact of the nappe beneath the Vavouto peninsula
438 (Fig. 2b), the obliquity of the main shear zones with respect to the boundaries of the
439 serpentine sole is probably around 22° . This obliquity is consistent with an interpretation of
440 the shear zones as large C'-type shear bands (Berthé et al., 1979; Passchier and Trouw, 1996)
441 developed during strong tangential shear across the serpentine sole. Smaller equivalent
442 structures are visible in Figures 12c, 13b and 13c. Finally, it should be noted that flat-lying
443 antithetic shear planes also occur in the eastern part of the VA1 section (Figs. 2c and 12b). In
444 Figure 12b, they crosscut the main southwest-dipping shear fabric, however their relationship
445 with one of the major shear zones suggests that they may have developed contemporaneously.

446 5. Discussion

447 5.1. The distribution of deformation in the Koniambo Massif

448 Figure 14 summarizes the results of our structural analysis while Figure 15 illustrates the
449 possible relationships between the three structural levels identified in the Koniambo Massif.

450 The Upper structural level is characterized by a very dense network of fractures filled
451 with one or several types of serpentine. Antigorite and polygonal serpentine commonly form
452 slickenfibers along fault planes, which we used for analyzing fault slip data. The bulk
453 deformation associated with antigorite-bearing faults is characterized by a highly
454 constrictional ellipsoid with λ_1 lying horizontally along a $N102^\circ$ trend. The bulk deformation
455 associated with polygonal serpentine-bearing faults is characterized by an ellipsoid in the
456 flattening field with λ_3 lying subhorizontally along a $N144^\circ$ trend.

457 The Lower structural level coincides with the serpentine sole. As a rule in New
458 Caledonia (e.g., Avias, 1967; Leguéré, 1976; Cluzel et al., 2012), the serpentine sole of the
459 Koniambo Massif is intensely deformed. Breccias dominate while foliated serpentinites,
460 interpreted as mylonitic rocks, form a distinct basal layer at least 20 m thick. The sole has
461 experienced pervasive non-coaxial deformation with a top-to-SW sense of shear. Polygonal
462 serpentine and magnesite are syn-kinematic phases with respect to this deformation. Along
463 the VA1 section, major low-dipping shear zones are distributed with a ~100 m lengthscale
464 (Fig. 2c). As described in Section 4.3.3, their mean obliquity with respect to the basal contact
465 of the nappe is estimated at ~22°. We interpret these shear zones as C'-type shear bands
466 developed within a thick zone of strong tangential shear that coincides with the serpentine
467 sole (Figs. 14 and 15; see also Lahondère et al., 2012, their figure 115). In line with this
468 interpretation, no equivalent shear zone is observed above the serpentine sole.

469 The Intermediate structural level is characterized by the presence of several meters-
470 thick reverse-slip shear zones. Top-to-SW shear zones are synthetic to pervasive shearing
471 along the serpentine sole and, likewise, involve polygonal serpentine and magnesite as syn-
472 kinematic mineral phases. Therefore, although we had no opportunity to observe this contact
473 in the field, we suppose that the shear zones root along the roof of the serpentine sole, the
474 whole sole acting as a décollement (Fig. 14 and, in Figure 15, geometric relationship 'r1'). In
475 Figure 15, the roof of the sole is schematically shown as a distinct fault; in reality, it is more
476 likely to coincide with a zone across which the intensity of brecciation progressively
477 diminishes. In addition, some shear zones may root at deeper level (Fig. 15, relationship 'r2'),
478 for instance along the relatively sharp contact between the foliated serpentinites and the
479 overlying serpentinite breccias (Fig. 3b) and/or along the basal contact of the nappe. This
480 possibility is suggested by the presence of a relatively large top-to-SW reverse-slip shear zone
481 at the northern end of the VA2 section (Fig. 3a) and by the characteristics of the fault zone of
482 Figure 11 (see Section 4.3.2 and below).

483 The NE-SW horizontal spacing between two consecutive top-to-SW shear zones, SZ1
484 and SZ2, is ~450 m. This is slightly less than the thickness of the Upper structural level in the
485 Koniambo Massif (Fig. 2b) and much less than the thickness of the Peridotite Nappe in the
486 Massif du Sud (≥ 1.5 km). This narrow spacing suggests that at least some of the shear zones
487 do not cross the nappe up to the surface but remain confined to the Intermediate structural
488 level. In fact, although rock exposures are abundant in the open pits of the top of the
489 Koniambo Massif, we have not observed any such shear zone in the Upper structural level.
490 Based on this indirect evidence, it is here suggested that the shear zones may connect to
491 another flat-lying décollement located along the roof of the Intermediate structural level (Fig.
492 15, relationship 'r3'). Like for the roof of the sole, this décollement might not coincide with a
493 distinct fault but with a zone across which the reverse-slip displacement of the shear zones
494 could be absorbed through diffuse faulting. In addition, a few major shear zones may ramp up
495 to the surface (Fig. 15, relationship 'r4'). No such shear zone is observed in the Koniambo
496 Massif but one likely exists in the northern part of the nearby Kopéto-Boulinda Massif (see
497 Figure 1 for location). It is highlighted by a north-dipping serpentinite sheet about 200 m
498 thick that emanates from the serpentine sole of this massif, at elevations around 200 m, and
499 climbs among the peridotites up to an elevation of at least 500 m (Maurizot et al., 1985, their
500 figure 4; Maurizot, 2007). The precise geometry and the kinematics of this probable shear
501 zone remain to be documented.

502 Finally, top-to-NE shear zones also exist in the Koniambo Massif, as subsidiary
503 structures in the serpentine sole and as conjugate shears with respect to the top-to-SW shear
504 zones in the Intermediate structural level. Because it roots along or slightly beneath the base
505 of the serpentine sole and likely offsets its roof (Fig. 2b), the reverse fault zone of Figure 11 is
506 probably a relatively late feature with respect to pervasive top-to-SW shearing along the sole.
507 The shear zone labelled 'r2', in Figure 15, would be an equivalent synthetic shear zone.

508 5.2. The temporal evolution of deformation

509 Our analysis documents major variations in the orientation of the principal strain axes. At
510 least part of these variations reflect a temporal evolution (Fig. 14). This is the case within the
511 Upper structural level, where fault slip data associated with antigorite and polygonal
512 serpentine were collected in the same restricted area but yielded two strikingly different strain
513 ellipsoids. Microstructural observations indicate that polygonal serpentine postdates antigorite
514 (see Sections 4.1 and 4.3.1), in agreement with the time sequence identified by Ulrich (2010)
515 in the serpentine sole of the Koniambo Massif. Hence, the Upper structural level keeps the
516 record of a temporal change from WNW-ESE horizontal stretching, during antigorite
517 crystallization, to NW-SE horizontal shortening, during polygonal serpentine crystallization.

518 The origin of the other variations in orientation of the principal strain axes is more
519 difficult to assess because they involve the same serpentine polymorph (polygonal serpentine)
520 and occur across distinct levels of the nappe. Therefore, the change from NW-SE shortening
521 in the Upper structural level to NE-SW shearing in the serpentine sole could reflect a spatial
522 rather than a temporal evolution, i.e. vertical strain partitioning at a specific evolutionary
523 stage of the nappe. The ~N-S direction of shortening recorded at an intermediate structural
524 height, on roadside '620 m', may support this view. Nevertheless, the observed changes in
525 shortening direction are more likely to reflect a temporal evolution. A first reason is that it
526 seems difficult to conceive a tectonic setting in which the nappe would undergo horizontal
527 shortening at right angle to its direction of displacement (assuming the latter is given by the
528 direction of shear along the basal décollement). A second reason is the fact that, in the
529 Intermediate structural level, the main shear zones, which accommodate NE-SW shortening,
530 crosscut rocks with the same dense network of fractures as in the Upper structural level (Fig.
531 10a). This relation suggests that the event having produced NE-SW shortening in the
532 Intermediate structural level (and top-to-SW shearing in the sole) is younger than the fracture

533 sets of the Upper structural level.

534 Thus, we suspect that both the syn-antigorite and the syn-polygonal serpentine fault
535 sets identified in the Upper structural level are older than the top-to-SW shear deformation
536 observed in the serpentine sole. This contrasts with the opinion of Leguéré (1976) who
537 supposed that the various sets of fractures he identified in the main mass of the nappe are all
538 younger than the deformation recorded by the sole. According to Leguéré (1976), these
539 fracture sets provide evidence for a three-step temporal evolution involving an episode of
540 ENE-WSW compression, then an episode of NNW-SSE compression, then an episode of
541 NNW-SSE extension. This scenario applies to the Kopéto-Boulinda and Koniambo Massifs,
542 but the same events and the same chronology, with some variations in the orientation of the
543 strain axes, are given by Leguéré (1976) for the other klippen of the Peridotite Nappe he
544 examined, spread over the Grande Terre. The episode of NNW-SSE compression, of fairly
545 constant orientation across the Grande Terre (Leguéré, 1976), correlates well with the syn-
546 polygonal serpentine event we identified in the Upper structural level of the Koniambo
547 Massif. The episode of NNW-SSE extension, which is more variable in orientation on the
548 scale of the Grande Terre (extension is locally WNW-ESE to W-E, cf. Leguéré, 1976), may
549 correlate with the syn-antigorite event we identified. Finally, the episode of ENE-WSW
550 compression, locally modified to a NE-SW compression (Leguéré, 1976), correlates with the
551 syn-polygonal serpentine±magnesite event we identified in the Intermediate structural level
552 and the serpentine sole. If so, however, the chronology of events proposed by Leguéré (1976)
553 differs strikingly from ours: his first event is our last one, and his last event is our first one.
554 We ignore the reasons for this disagreement, but note that Leguéré (1976) did not provide
555 details on the arguments he used for establishing his chronology. Our chronology is
556 essentially based on the identification of a sequence of minerals associated with deformation,
557 from antigorite to polygonal serpentine±magnesite. This sequence is consistent with a
558 progressive lowering in temperature conditions (Ulrich, 2010), which may reflect a

559 modification of the geothermal gradient and/or the progressive reduction in thickness of the
560 nappe (Fig. 15). As a possible explanation for our disagreement with Leguéré (1976), we
561 might have failed to identify one or several late episodes of normal faulting that are reported
562 from the sedimentary series surrounding the klippe of the Peridotite Nappe (Lagabrielle et
563 al., 2005; Chardon and Chevillotte, 2006).

564 5.3. Relationships between deformation and serpentinization

565 The three structural levels identified in the Koniambo Massif show strikingly different styles
566 of deformation. These levels correlate fairly well with the subhorizontal lithological layering
567 of the massif defined by Maurizot et al. (2002), which refers to variations in the intensity of
568 serpentinization. Hence, it is tempting to consider that the intensity of serpentinization played
569 a major role in the way deformation has been distributed across the Peridotite Nappe. As
570 implicitly considered in several studies (e.g., Avias, 1967; Cluzel et al., 2001), the massive
571 serpentinites of the sole might represent a layer of soft rocks having led to the
572 individualization of a major décollement at the base of the nappe.

573 The Upper structural level displays a dense network of serpentine-bearing fractures,
574 many of which have been activated as faults. Nevertheless, the bulk deformation associated
575 with these faults is probably limited, as suggested by the fairly constant orientation of dunitic
576 interlayers all across the Upper structural level (Maurizot et al., 2002). Moreover, as
577 discussed in Section 5.2, the deformation recorded by the serpentine sole and the shear zones
578 of the Intermediate structural level is likely younger than most fault sets of the Upper
579 structural level. This supports the picture of an almost rigid Upper structural level by the time
580 shearing was accumulating along the sole (cf. the strain-depth diagram in Figure 15). Could
581 this behaviour relate to a least degree of bulk serpentinization in the Upper structural level?

582 Unlike in some parts of the Massif du Sud (e.g., Orloff, 1968), pristine peridotites are
583 virtually absent from the Koniambo Massif. The least serpentinized rocks are found in the

584 highest ~200 m of the massif. In the field, they are identified as they seem to retain a
585 granoblastic texture, yet their dark green colour is symptomatic of a significant serpentine
586 content. Under microscope, these samples show ‘mesh’ textures resulting from a partial
587 replacement of olivine by lizardite (Fig. 6b). Serpentine mesh textures are essentially
588 pseudomorphic, therefore peridotites with these textures usually preserve their pre-
589 serpentinization fabrics (e.g., Wicks and Whittaker, 1977; Rouméjon and Cannat, 2014),
590 which also applies in that case (Fig. 6b; for other examples from the Koniambo Massif, see
591 Lahondère et al., 2012, their figures 301 to 304 and 309 to 312). This suggests that the Upper
592 structural level behaved almost rigidly ever since the early development of lizardite.

593 The mesh texture in Figure 6b is classical (see also the examples of Lahondère et al.,
594 2012) in the sense that lizardite forms connected ‘mesh rims’ surrounding isolated olivine
595 ‘mesh cores’ (e.g., Rouméjon and Cannat, 2014). This is a key observation because
596 theoretical considerations (e.g., Handy et al., 1999) as well as experimental results (Escartín et
597 al., 2001) indicate that, with such a texture, the peridotite should be nearly as weak as pure
598 serpentine. The rock in Figure 6b contains ~25% serpentine (estimated from image analysis)
599 whereas experimental evidence suggests that a ~10-15% serpentine content is enough for
600 achieving the largest part of the strength drop with respect to the strength of a pure peridotite
601 (Escartín et al., 2001). Therefore, it seems that the highly uneven vertical distribution of syn
602 to post-serpentinization deformation across the height of the Koniambo Massif does not relate
603 to a parallel gradient in rock strength: within the range of observed variations in the degree of
604 serpentinization, all the rocks should have approximately the same strength (cf. Escartín et al.,
605 2001).

606 5.4. What promoted strain localization at lower levels of the nappe ?

607 The above discussion has shown that the degree of serpentinization is probably not the factor
608 which led deformation to localize along the sole of the nappe. A simple alternative may be

609 considered where the gradient in rock strength is not located within the Peridotite Nappe but
610 arises from its juxtaposition against rocks of the substratum. In the Koniambo Massif, as for
611 the other klippe of the northwestern coast, this substratum is essentially composed of basalts
612 and dolerites (e.g., Guillon, 1975; Paris, 1981; Cluzel et al., 2001) (Fig. 3a). Such rocks are
613 clearly stronger than serpentinites (e.g., Ildefonse et al., 2007) and, following Escartín et al.
614 (2001), also stronger than slightly serpentinized peridotites. Hence, a strength profile may be
615 conceived where the whole Peridotite Nappe is weak, overlying a stronger substratum.
616 Deformation is then expected to concentrate in the part of the weak domain that is the closest
617 to the strong one, i.e. on the soft side of the main rheological boundary. This could account
618 for strain localization along the sole of the nappe.

619 Where basalts and dolerites underlie the Peridotite Nappe, the thickness of the
620 serpentine sole is great (~100-200 m). This is the case in the Koniambo Massif and the other
621 klippe of the northwestern coast (e.g., Guillon, 1975). In contrast, smaller klippe located
622 further northeast, around the axis of the island, have a much thinner serpentine sole (~10-20
623 m) and a substratum made of pervasively schistose fine-grained sediments with a low grade
624 metamorphic overprint. Typical examples are the Tchingou Massif (e.g., Maurizot et al.,
625 1985) and a series of kilometer-sized klippe around the Ougne summit (Maurizot et al.,
626 1989) (Fig. 1). Assuming that the thickness of the sole depends essentially on the intensity of
627 shearing, and assuming that the total amount of shear, linked to the displacement of the nappe,
628 is approximately constant over the area under consideration, this large difference in sole
629 thickness is consistent with the hypothesis that a lower vs. higher amount of strain is
630 accommodated by the nappe when its substratum is made of soft metasediments vs. strong
631 mafic rocks, respectively. This supports our proposal that strain localization along the base of
632 the nappe results from its juxtaposition against a stronger substratum.

633 **6. Conclusion**

634 Three structural levels have been identified in the Koniambo Massif. The Upper structural
635 level, characterized by a dense network of fractures, keeps the record of at least two
636 deformation events, the first associated with antigorite (WNW-ESE extension), the second
637 with polygonal serpentine (NW-SE compression). The Lower structural level is represented
638 by the serpentine sole. It consists of massive tectonic breccias overlying a layer of mylonitic
639 serpentinites. The sole, which records pervasive tangential shear with top-to-SW kinematics,
640 can be interpreted as a décollement at the base of the nappe. The Intermediate structural level
641 exposes several meters-thick conjugate shear zones accommodating NE-SW shortening. Like
642 the sole, these shear zones involve polygonal serpentine and magnesite as the main syn-
643 kinematic mineral phases. The shear zones likely root into the basal décollement, either along
644 its roof or, occasionally, around its base.

645 Even the least altered peridotites, in the Upper structural level, contain so much
646 serpentine that, according to theoretical and experimental work, they should be nearly as weak
647 as pure serpentinite. Hence, no strong vertical gradient in strength due to variations in the
648 degree of serpentinization is expected within the exposed part of the nappe. Strain localization
649 along the serpentine sole probably results from the juxtaposition of the nappe, made of weak
650 serpentinized peridotites, against stronger mafic rocks. In line with this view, the serpentine
651 sole is about ten times thinner when the substratum of the nappe is made of weak
652 metasediments, compared to what it is when the substratum is made of basalts and dolerites.

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658

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827 **Figure captions**

828 Fig. 1. Simplified geological map of the Grande Terre, New Caledonia, showing the
829 exposures of ultramafic rocks. Most of these rocks are originally part of the Peridotite Nappe.
830 Adapted from Maurizot and Vendé-Leclerc (2009).

831 Fig. 2. a) Geological map of the central and southern parts of the Koniambo Massif, adapted
832 from Carroué (1972) and Maurizot et al. (2002). b) General cross-section of the Koniambo
833 Massif, located in (a). c) Report of the main fault zones along the VA1 section, in the
834 Vavouto peninsula.

835 Fig. 3. Two field views at the (a) northern and (b) southern ends of the VA2 section, in the
836 Vavouto peninsula (location in Figure 2a).

837 Fig. 4. Close views of the (a) first, (b) second and (c & d) third reference samples discussed in
838 the text, and corresponding Raman spectra.

839 Fig. 5. Field views of serpentine-bearing (a to d) fault planes and (e & f) shear zones in the
840 Upper structural level.

841 Fig. 6. a) Field view of an open pit with a dense network of fractures, typical from the Upper
842 structural level (location in Figure 2a). b) Photomicrograph of a sample representative of the
843 least serpentinized peridotites within the Koniambo Massif, from the Upper structural level, in
844 between the macroscopic fractures. The sample shows a typical 'mesh' texture composed of
845 lizardite (Liz) mesh rims surrounding olivine (Ol) and orthopyroxene (Opx) mesh cores.

846 Fig. 7. Lower hemisphere, equal-area projection showing the orientation of all measured (a)
847 syn-antigorite and (b) syn-polygonal serpentine faults at higher levels of the Koniambo
848 Massif (location in Figure 2a), and (c & d) the result of the analysis of these fault slip data
849 using the FaultKin software of Allmendinger et al. (2012). Kamb contours are shown for the
850 axes of maximum stretching (in c) or maximum shortening (in d) with intervals in shades of
851 grey as depicted in the legend. The parameter R equates $(\epsilon_2 - \epsilon_3) / (\epsilon_1 - \epsilon_3)$ and describes the

852 shape of the strain ellipsoid (R is 0 for pure constriction, 0.5 for plane strain, and 1.0 for pure
853 flattening). N_{max} is the number of faults compatible with the computed ellipsoid.

854 Fig. 8. Lower hemisphere, equal-area projection showing the orientation of faults at two
855 specific sites (location in Figure 2a), and the result of the analysis of these fault slip data.

856 Fig. 9. Lower hemisphere, equal-area projection showing the orientation of faults and minor
857 shear zones at roadsite '620 m' (location in Figure 2a), and the result of the analysis of this
858 fault slip dataset. Kamb contours are shown for the axes of maximum shortening with
859 intervals in shades of grey as depicted in the legend.

860 Fig. 10. Field views of (a to d) SZ1 and (e) SZ2, two major low-dipping shear zones typical
861 from the Intermediate structural level (location in Figure 2a). Plg, polygonal serpentine; Mgs,
862 magnesite. f) Lower hemisphere, equal-area projection showing the orientation of structural
863 elements in the two shear zones (see the text).

864 Fig. 11. Field views of the reverse-slip fault zone with top-to-ENE kinematics shown in
865 Figure 2a,b.

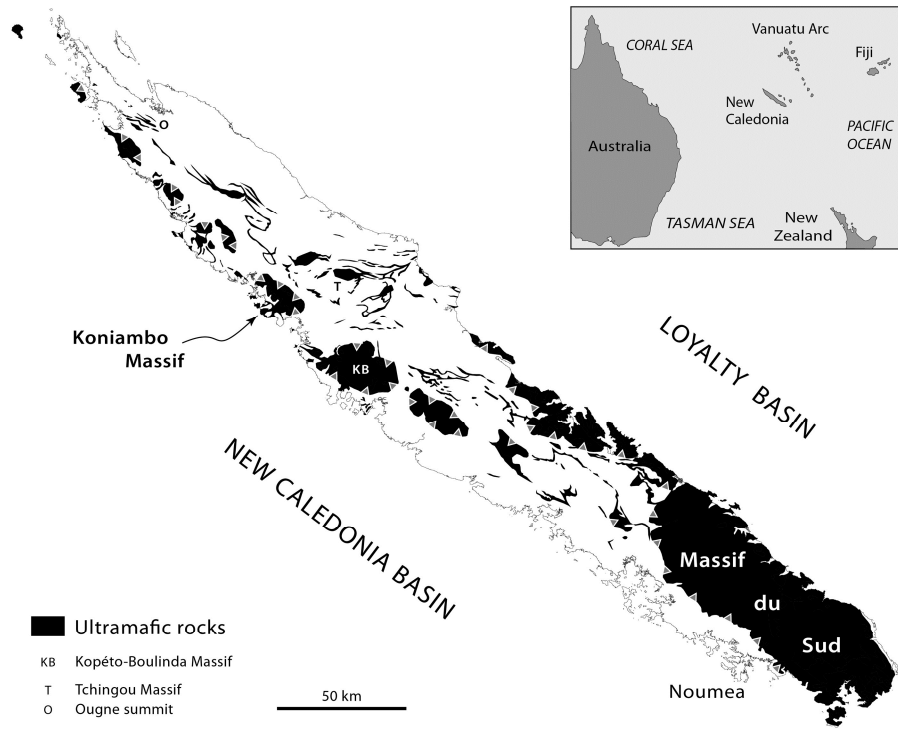
866 Fig. 12. Field views of the serpentine sole in the Vavouto peninsula, along the (a & b) VA1
867 and (c & d) VA2 sections (location in Figure 2). Plg, polygonal serpentine; Mgs, magnesite.
868 e) Lower hemisphere, equal-area projection showing the orientation, along the VA1 section,
869 of large shear planes (shown as orange squares representing poles of planes), of planar
870 magnesite veins (shown as great circles) in between major shear zones, and of fold axes
871 (shown as black dots) of folded magnesite veins within the shear zones.

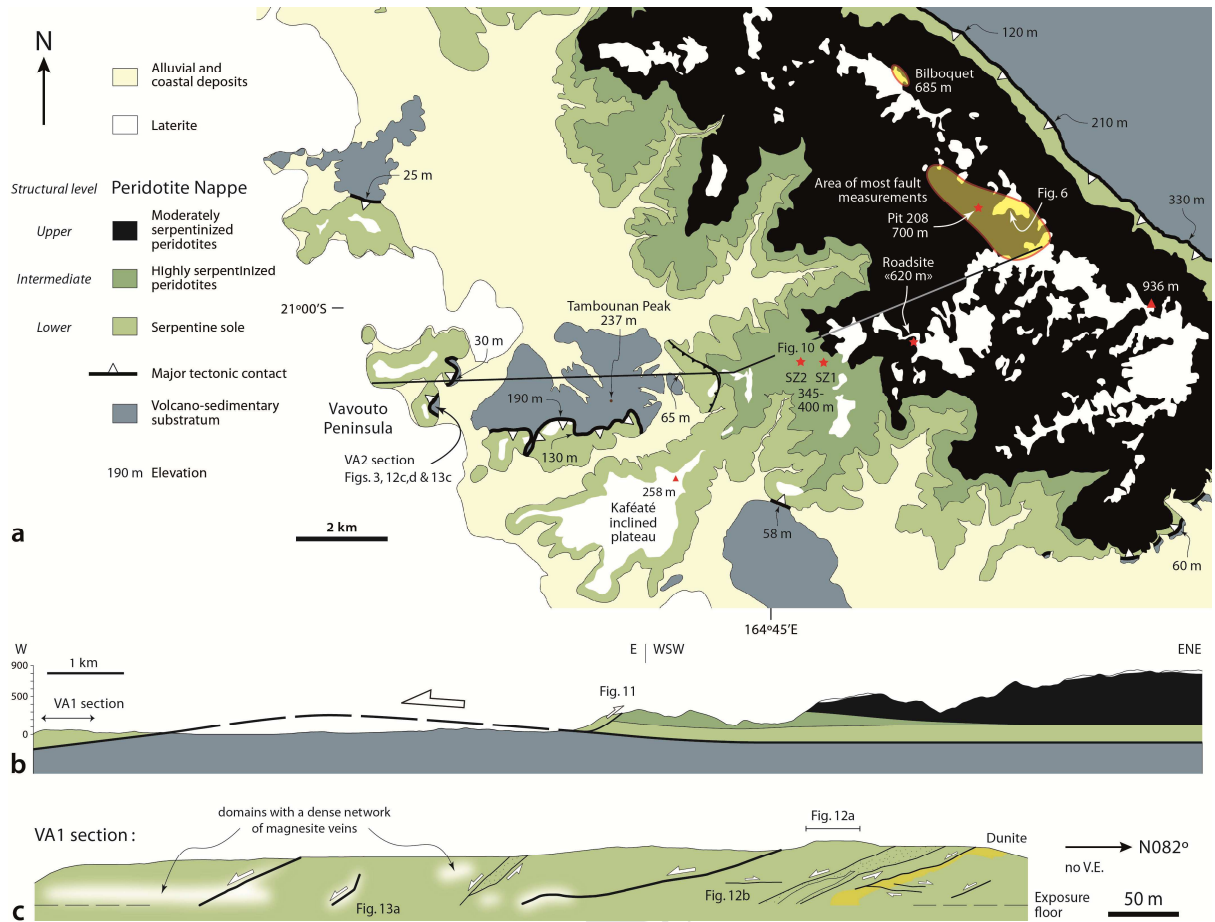
872 Fig. 13. Field views (a to c) and microphotograph (d) illustrating the relationships between
873 magnesite and deformation in the serpentine sole.

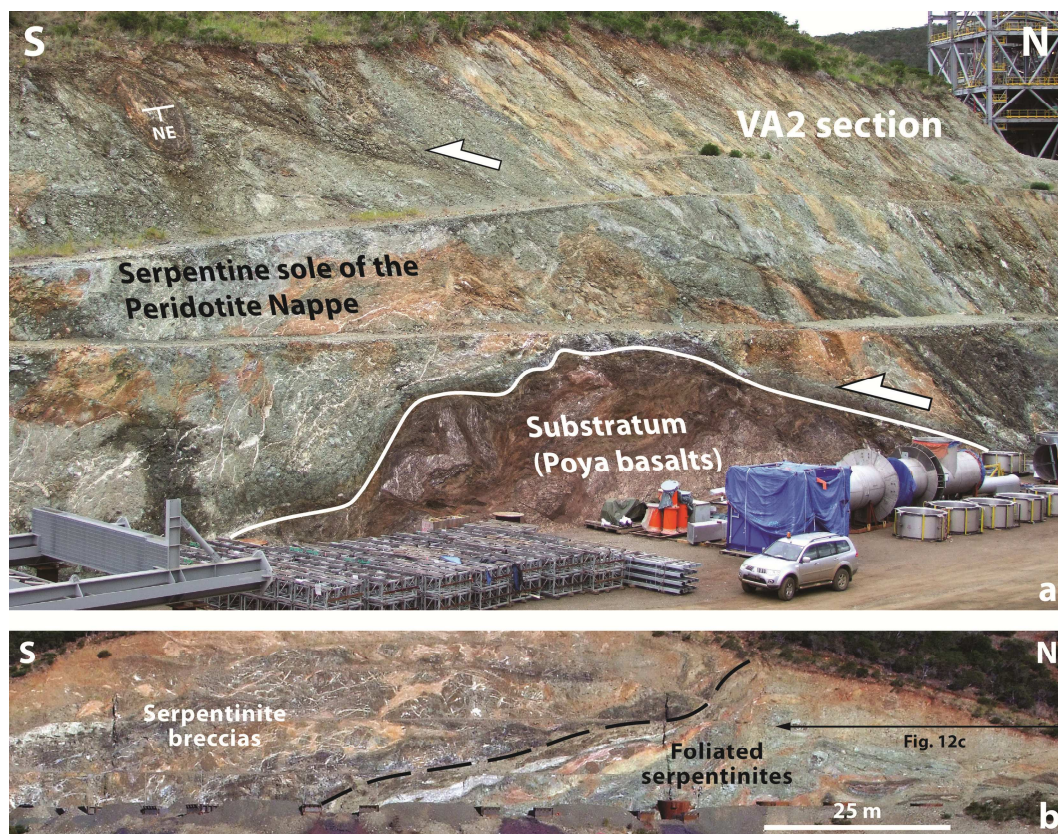
874 Fig. 14. Left, block-diagram depicting the style of deformation in the three structural levels
875 identified in the Koniambo Massif. For the Upper structural level, the drawing shows the

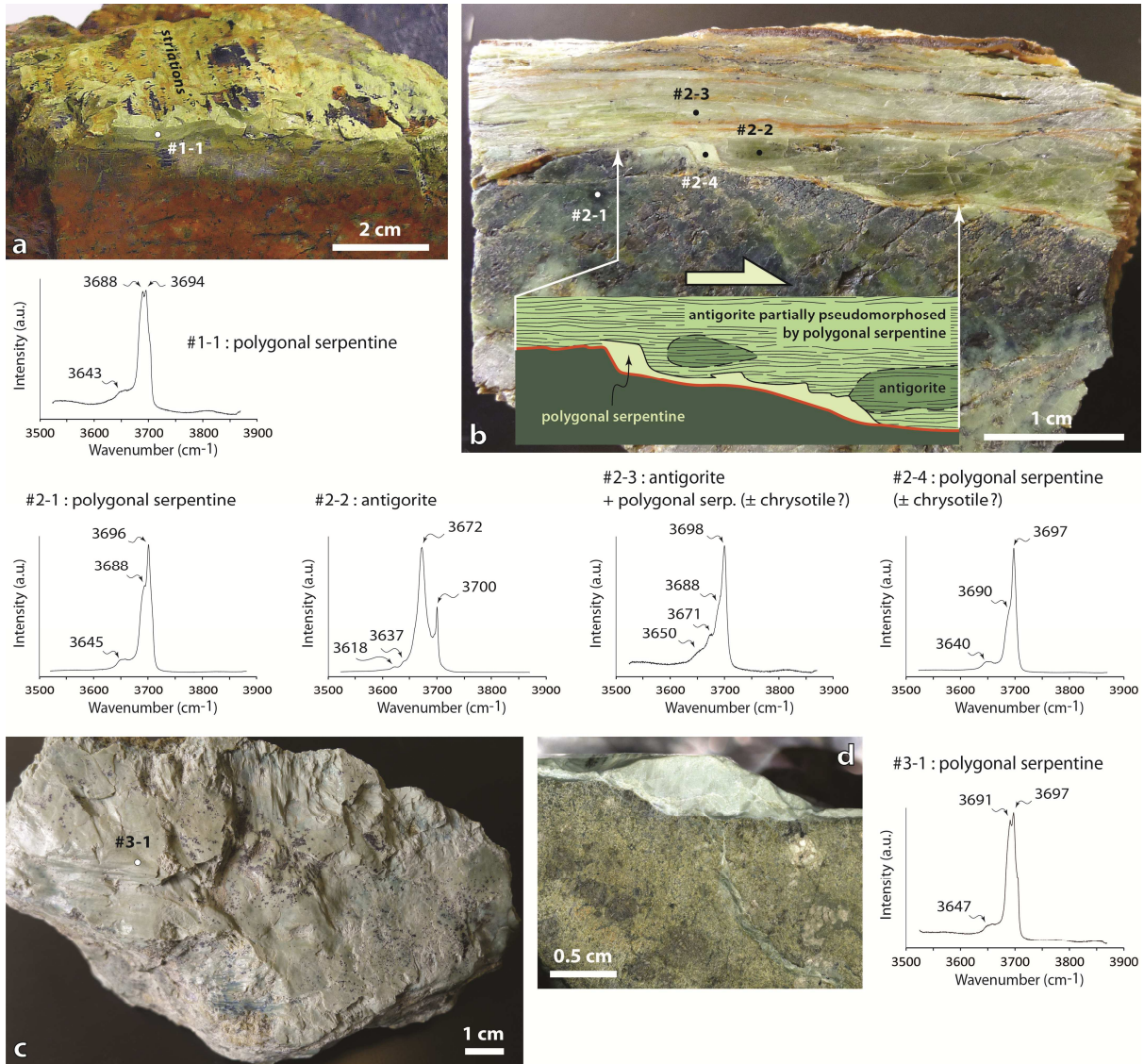
876 deformation associated with polygonal serpentine, rather than with antigorite, because
877 polygonal serpentine is also the main mineral phase associated with major structures in the
878 other structural levels. Roadsite '620 m' is shown at the top of the Intermediate structural
879 level to highlight its location at lower elevation in comparison to the main area of fault
880 measurements (see Figure 2a). Right, synthesis of the results of strain inversion as a function
881 of the structural level (vertically) and as a function of the main minerals associated with
882 deformation, which enable the determination of a sequence of events (horizontally).

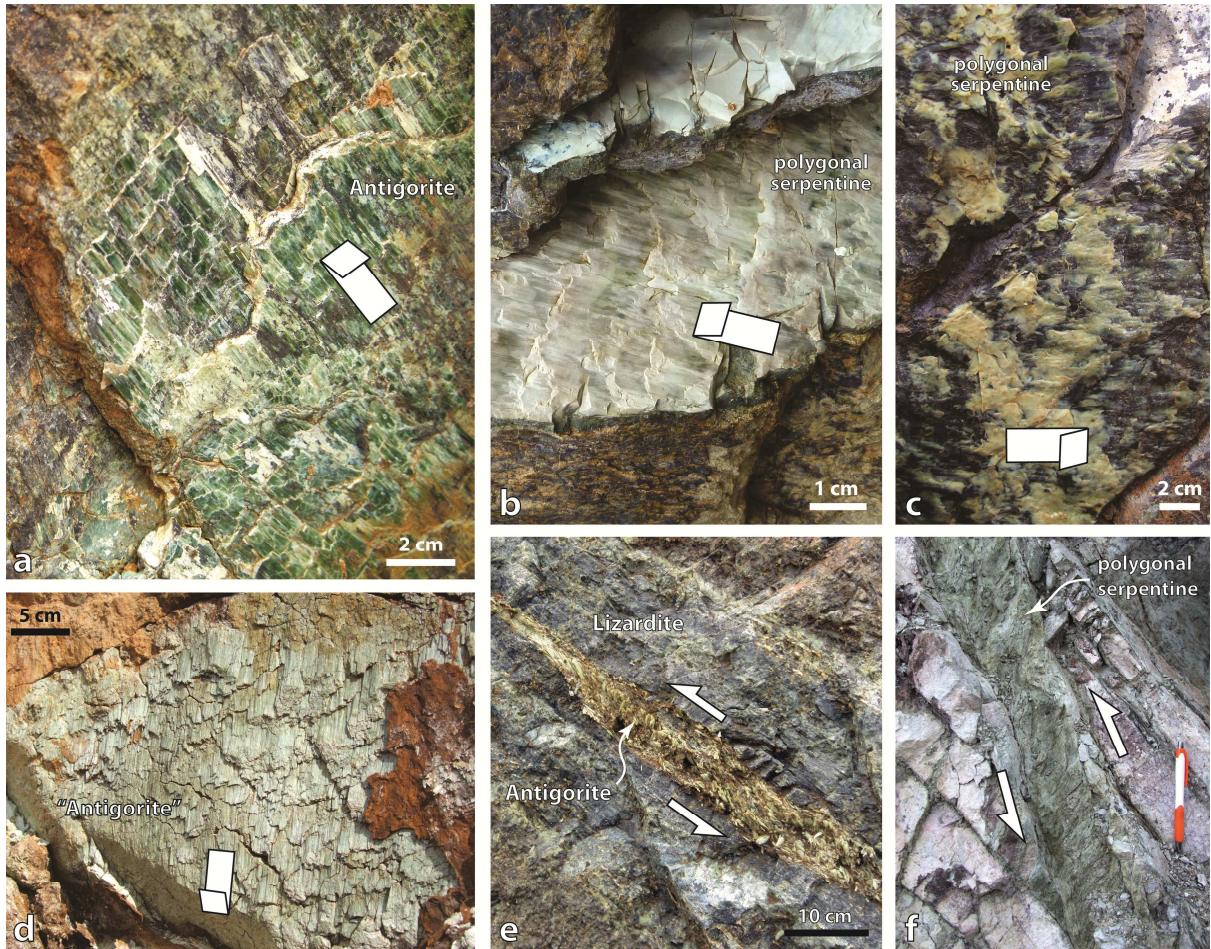
883 Fig. 15. Schematic cross-section depicting the vertical distribution of deformation within the
884 Peridotite Nappe by the time top-to-SW shearing occurred along the serpentine sole. 'r1' to
885 'r4' are geometric relationships discussed in the text. The thickness and the topography of the
886 nappe are poorly constrained and are likely to have changed during deformation.

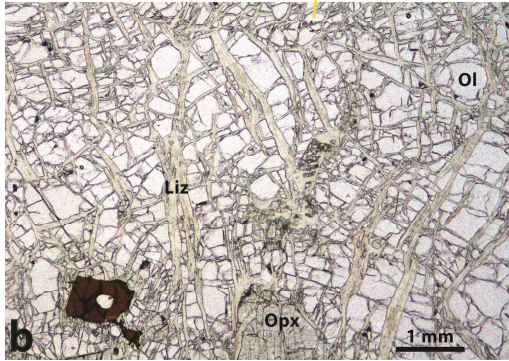
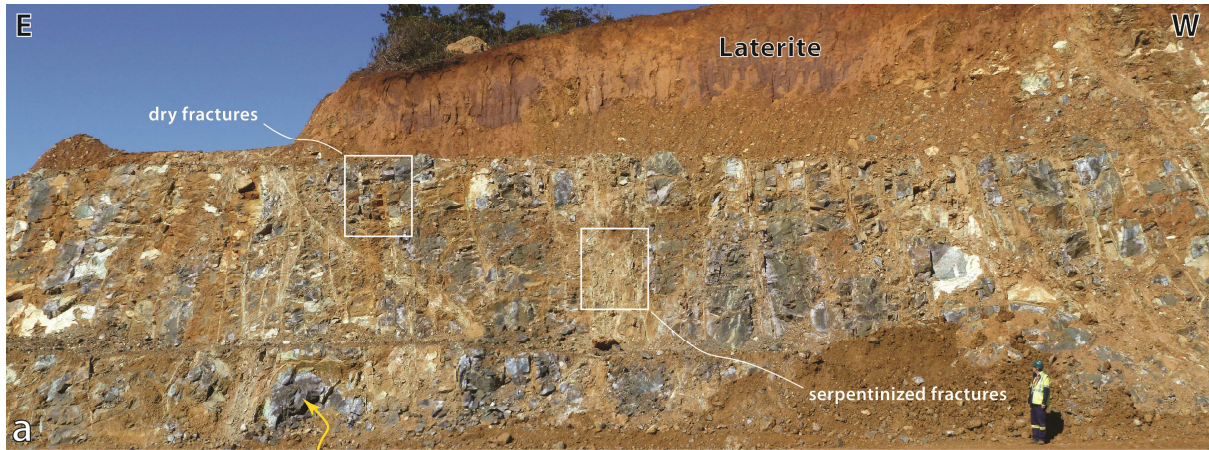


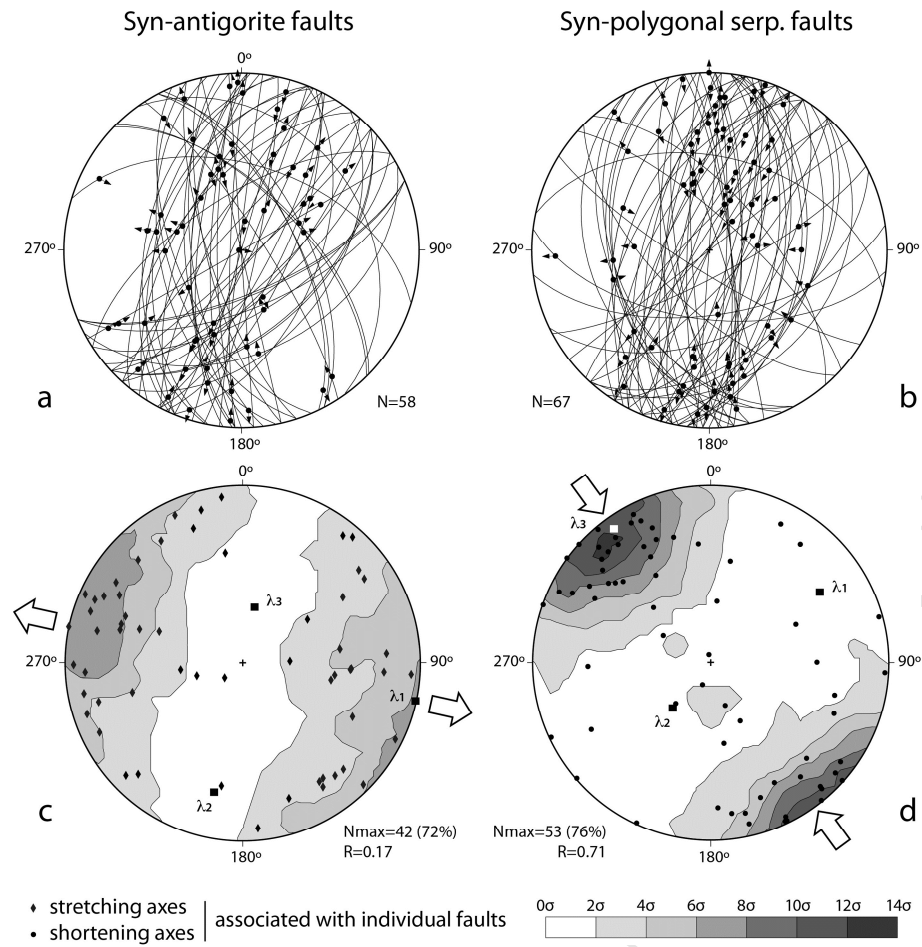




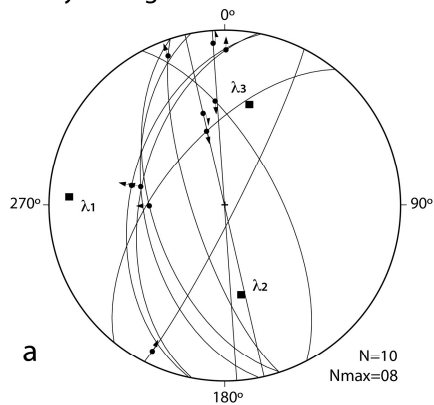




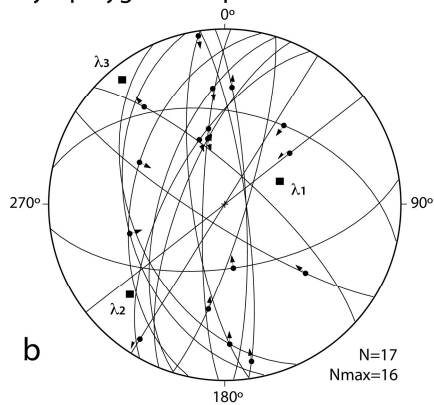




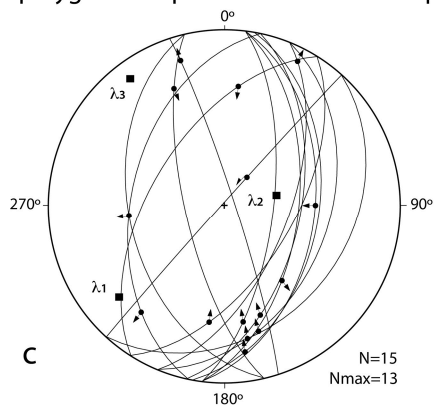
Syn-antigorite faults on Pit 208



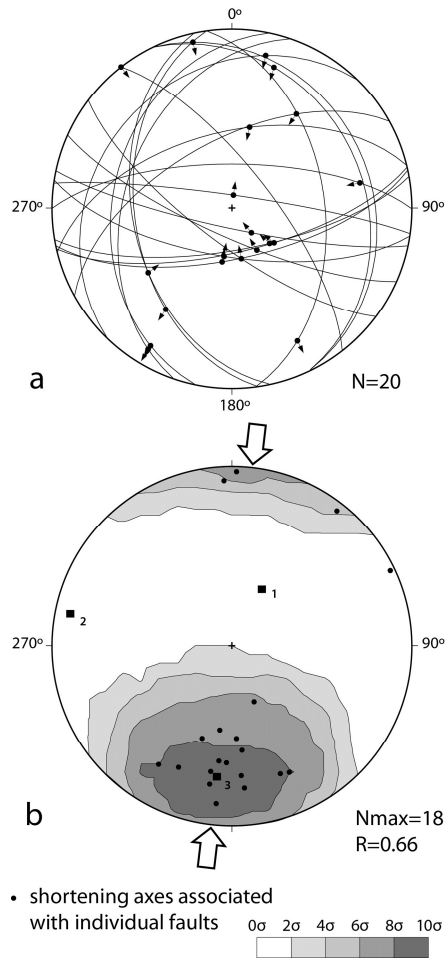
Syn-polygonal serp. faults on Pit 208

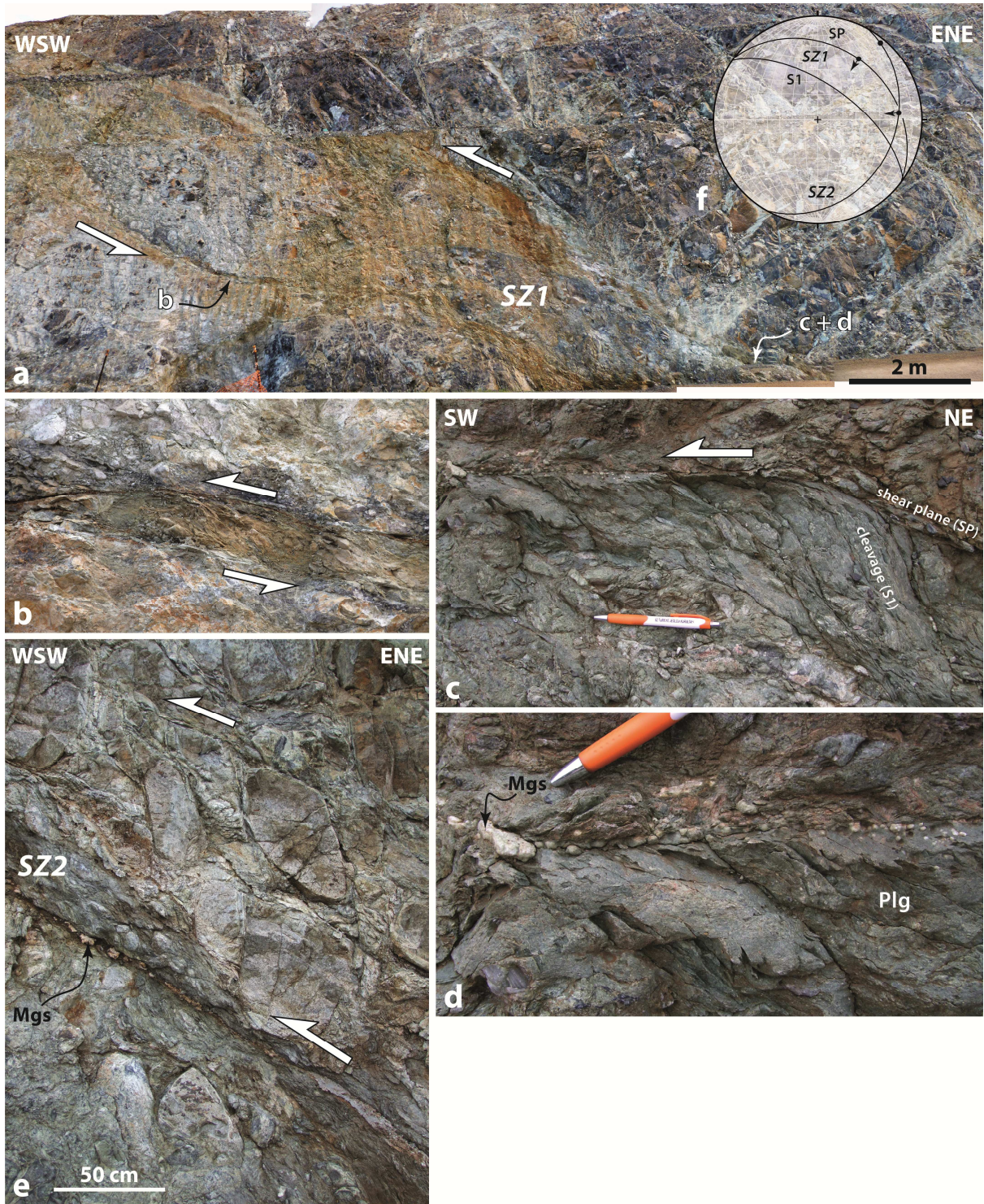


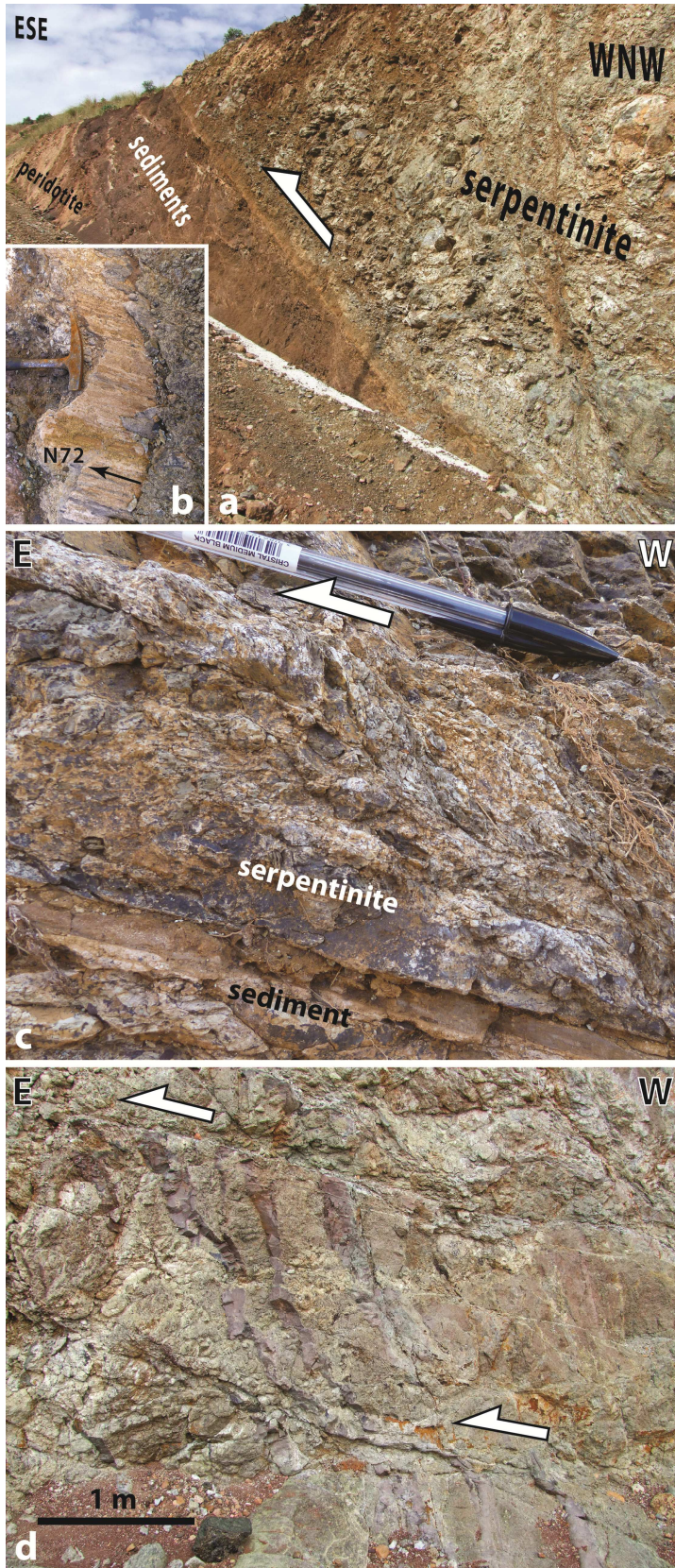
Syn-polygonal serp. faults on site «Bilboquet»

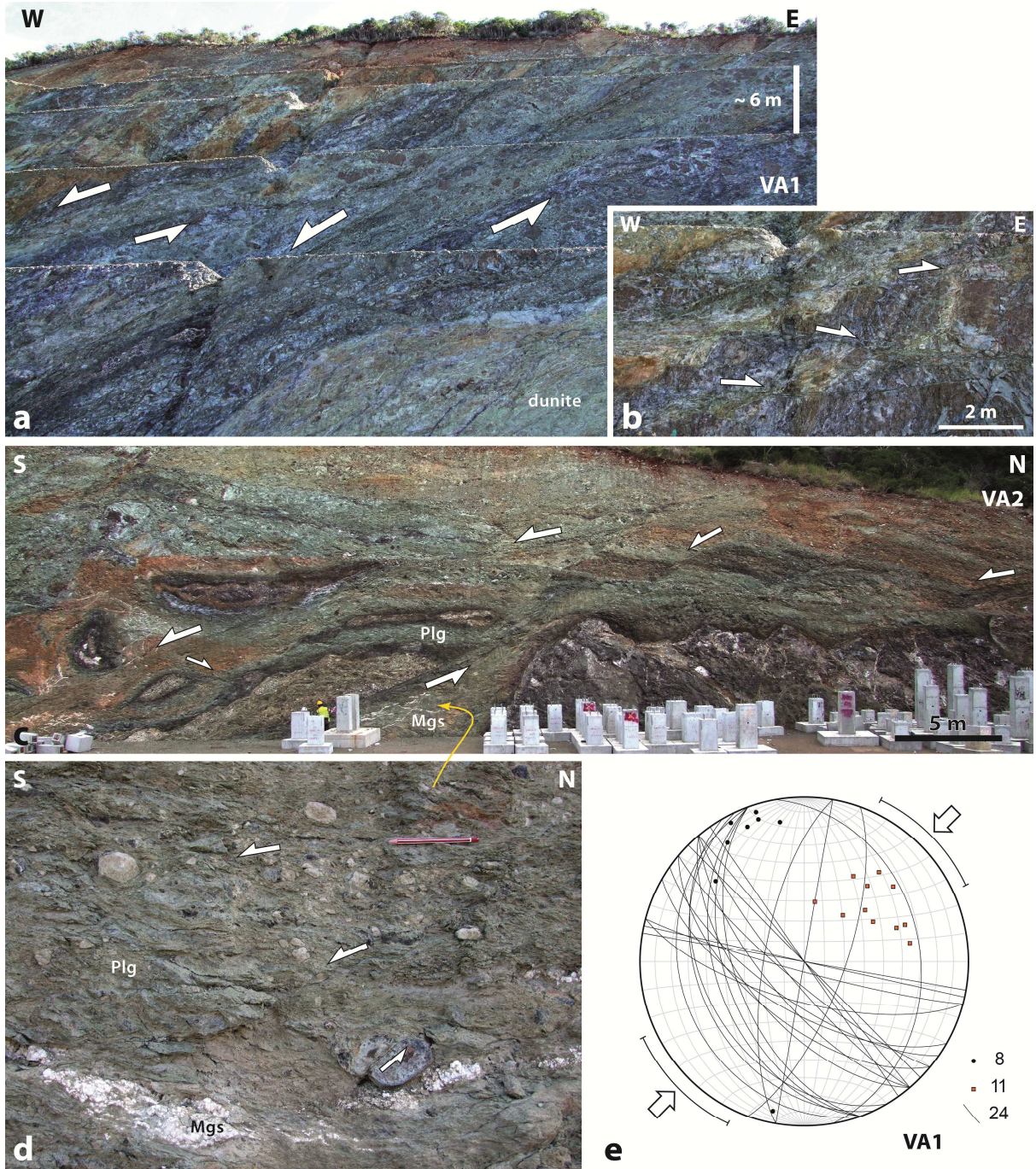


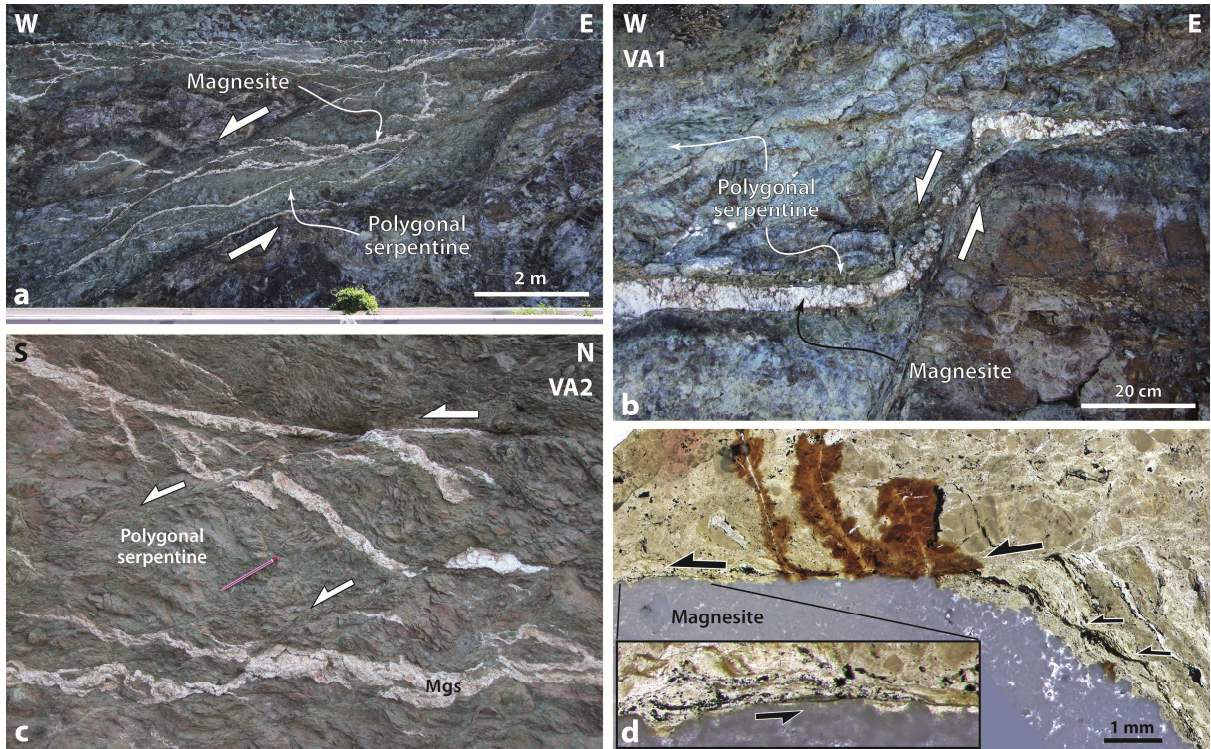
Syn-polygonal serpentine faults
and minor shear zones,
roadside «620 m»

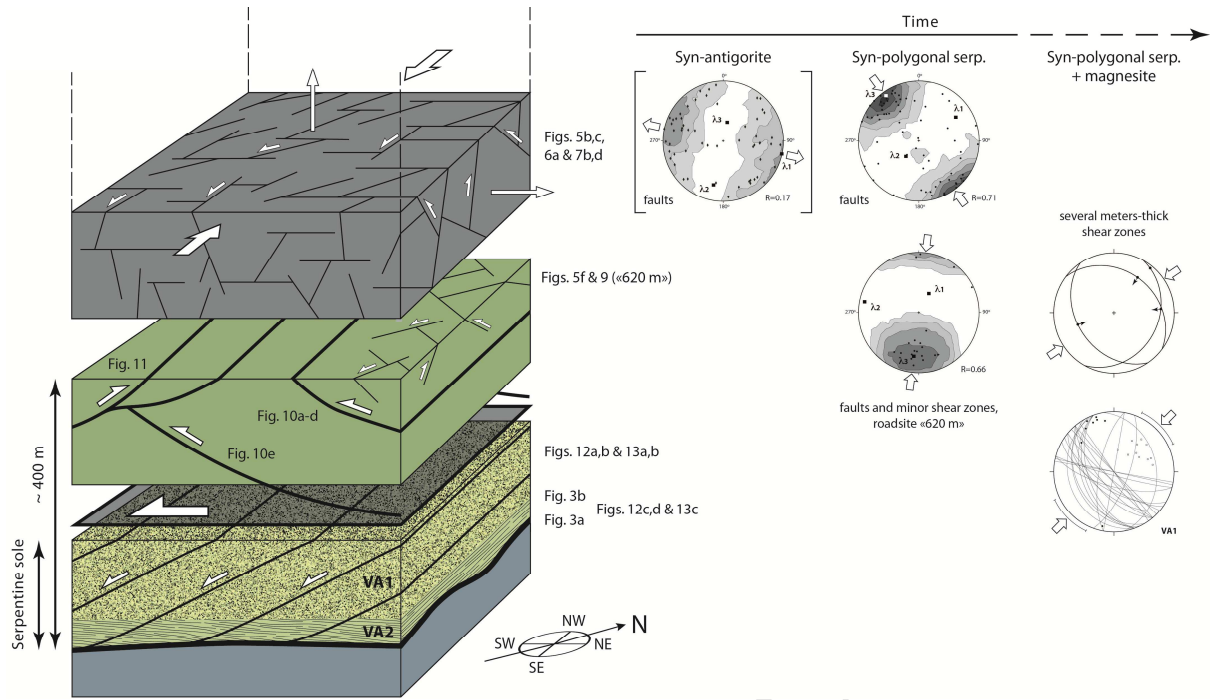




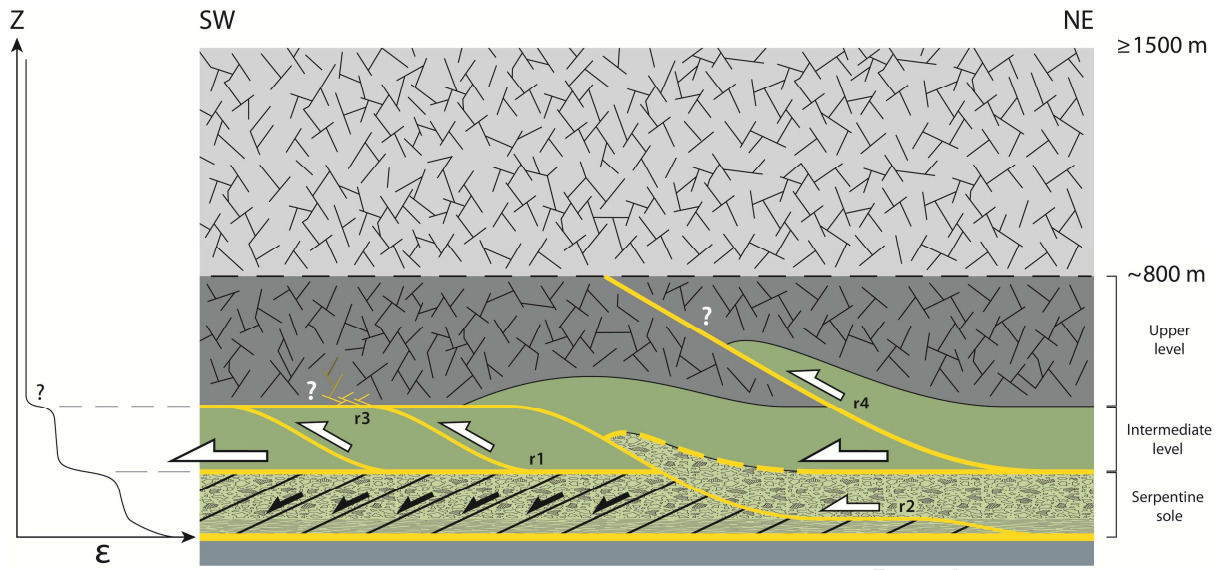








ACCEPTED MANUSCRIPT



- We study the deformation associated with serpentines in the New Caledonia ophiolite
- Structural analysis is coupled with mineralogical characterization
- The evolution of deformation through space and time is described
- The reason for strain localization along the sole of the nappe is discussed

ACCEPTED MANUSCRIPT