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Serpentinites of Different Tectonic Origin in an Exhumed Subduction Complex (New Caledonia, SW Pacific)

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Key Points:

- Two compositionally-distinct types of serpentinites identified in the New Caledonia subduction complex

- Some serpentinites derive from the subducting plate; others from the overlying mantle wedge

- New Caledonia preserves evidence for incorporation and exhumation of mantle hangingwall via the subduction channel

22   **Abstract**

23   Owing to the importance of serpentinites for planetary geochemical and geodynamic

24   processes, there has been much work discerning the origins of their parent rocks, including

25   distinguishing between serpentinites derived from a subducting plate vs. overlying mantle in

26   exhumed subduction complexes. The island of New Caledonia (SW Pacific Ocean) provides a

27   rare window into Cenozoic Pacific subduction processes. The island is unique in exposing both

28   an exceptionally-preserved high-pressure, low-temperature subduction complex and one of the

29   largest supra-subduction zone ophiolites in the world. Previous studies disagree on the origin of

30   serpentinites in the subduction complex. In this study, we analyze twenty-three serpentinites

31   from this subduction complex for whole-rock major and trace element geochemistry and stable

32   isotope (δD, $\delta^{18}O$) compositions. Our data reveal two distinct groups of serpentinites: Group I

33   samples in the northern portion of the complex are pervasively serpentinized, and exhibit

34   enriched heavy rare earth element (REE) compositions and $\delta^{18}O$ between +6.7‰ and 10.2‰. In

35   contrast, Group II serpentinites in the south preserve relict orthopyroxene and olivine, and show

36   depleted trace element compositions and comparatively lower $\delta^{18}O$ values between +5.1‰ and

37   +8.0‰. We interpret Group I serpentinites to derive from downgoing plate mantle, whereas

38   Group II serpentinites derive from overlying mantle wedge, exhibiting remarkable similarity to

39   the REE geochemistry of the structurally-overlying New Caledonia ophiolite. Our results

40   establish the subduction complex in New Caledonia as an unusual natural record of the

41   entrainment and exhumation of mantle from both the overlying mantle wedge and the

42   downgoing plate in an oceanic subduction zone.

43   **Plain Language Summary**

44   The hydration of Earth's mantle produces rocks called serpentinites that are important to

45   chemical cycling within the Earth system. This process, a form of metamorphism, occurs in

46   several types of tectonic settings on Earth. Serpentinites formed in these different settings are

47   imparted with unique geochemical "fingerprints" due to different types of fluids and varied

48   compositions of the original mantle material. In metamorphic complexes that preserve remnants

49   of subduction zones, serpentinites can offer clues to the original tectonic setting and subsequent

50   mechanics of subduction. We analyze stable isotope and whole-rock compositions to determine

51   the origin of the mantle that produced serpentinites found in the metamorphic complex preserved

52 in New Caledonia (SW Pacific). We discover two distinct groups of serpentinites: one group in

53 the northern portion of the complex likely formed on the ocean floor prior to subduction, and

54 experienced high degrees of fluid alteration as it was metamorphosed. In contrast, a second

55 group of serpentinites in the southern portion of the complex resembles material from the mantle

56 overlying the subducting plate and is less altered. This locality represents an uncommon

57 example, globally, where material from this overlying mantle is entrained and preserved at the

58 surface in an exhumed subduction complex.

59 **1 Introduction**

60   Serpentinites are hydrated fragments of Earth's mantle and commonly occur in exhumed

61 subduction complexes. They are important vehicles in the global cycling of water, carbon,

62 nitrogen, fluid-mobile elements (FME), and halogens, releasing these elements by prograde

63 dehydration at forearc to subarc depths during subduction (e.g., Alt et al., 2013; Barnes et al.,

64 2018; Collins et al., 2015; Deschamps et al., 2011; Halama et al., 2014; Hattori & Guillot, 2003;

65 John et al., 2011; Kerrick & Connolly, 1998; Kodolanyi et al., 2012; Scambelluri et al. 2004,

66 2019; Tenthorey & Hermann, 2004; van Keken et al., 2011). Serpentinites and their hybridized,

67 metasomatized derivatives formed at depth in subduction zones are critical in producing the

68 distinct chemical composition of arc magmas (e.g., Codillo et al. 2018; Hattori & Guillot, 2003;

69 Marschall & Schumacher, 2012; Nielsen & Marschall, 2017; Shimoda & Kogiso, 2019; Tatsumi,

70 1986) and contribute significantly to the net redox budget of subducted oceanic plates and the

71 oxidation state of dehydration fluids infiltrating the overlying mantle (Debret & Sverjensky,

72 2017; Evans, 2012; Evans & Frost, 2021; Evans et al., 2017). The wide stability field of the

73 serpentine mineral antigorite also makes possible the transport of these chemical components

74 deeper into the mantle, past subarc depths (e.g., Hacker, 2008; Hattori & Guillot, 2003; Kendrick

75 et al., 2011, 2017; Scambelluri & Tonarini, 2012; Schmidt & Poli, 1998; Ulmer & Tromsdorff,

76 1995; Wunder & Schreyer, 1997).

77   Serpentinites are also critical to the rheology of the subducting zone, and in particular the

78 plate interface (Gerya et al., 2002; Hermann et al., 2000; Rüpke et al., 2004; van Keken et al.,

79 2011). Their high water contents and low density facilitate the exhumation of denser blueschist

80 and eclogite (Guillot et al., 2000; Magott et al., 2020; Schwartz et al., 2001), and serpentinization

81 of the cold "nose" of the mantle wedge carries implications for mantle wedge flow and the

82 minimum depth of decoupling of the subducting plate and overlying mantle (Hilairet & Reynard,

83 2009; Kerswell et al., 2021; Reynard, 2013; Wada et al., 2008). Furthermore, serpentinites have

84 been correlated with zones of slow slip and intermediate-depth seismicity in subduction zones,

85 possibly associated with dehydration embrittlement resulting from the breakdown of antigorite

86 (Behr & Bürgmann, 2021; Ferrand, 2019; Hacker et al., 2003; Hilairet et al., 2006, 2007; Hirth &

87 Guillot, 2013; Jung & Green, 2004; Peacock, 2001; Proctor & Hirth, 2015; Toffol et al., 2022).

88 In short, the occurrence, composition, and spatial distribution of serpentinites and associated

89 metasomatic rocks in paleo-subduction complexes reveals information about first-order

90 subduction processes, including the transport and chemical evolution of fluids during subduction

91 and exhumation.

92   In exhumed subduction complexes, mantle material may originate from the downgoing

93 plate or the overriding plate (Figure 1). Prior to subduction, serpentinites form in the (eventual)

94 downgoing plate via hydrothermal circulation at ridge-transform systems where lithospheric

95 mantle underlying the igneous oceanic crust or exposed at/near the seafloor by extension is

96 hydrated by heated seawater (Barnes & O'Neil, 1969; Bonatti, 1976; Cannat, 1993; Cannat et al.,

97 2010; Mével, 2003; Rouméjon et al., 2015). Hydration may also occur as the plate approaches

98 the trench, if water circulates through slab-bend faults (Ranero et al., 2003). In the overriding

99 plate, the mantle wedge experiences hydration from fluid fluxing off the dehydrating downgoing

100 plate (e.g., Bostock et al., 2002; Fyfe & McBirney, 1975; Hyndman & Peacock, 2003).

101   Serpentinites formed in each of these distinct tectonic settings are geochemically distinct

102 owing to differences in (1) mantle protolith composition (i.e., melt depletion and refertilization

103 histories and degrees of fractional crystallization) and (2) the chemistry and nature of fluid

104 alteration, such as fluid-rock ratio and temperature of serpentinization (Figure 1) (e.g.,

105 Deschamps et al., 2010, 2013; Peters et al., 2017). These distinctions are evident in bulk-rock

106 major and trace element geochemistry, stable isotopes (e.g., O, H, Cl, B), and the composition of

107 relict primary minerals (e.g., spinel, pyroxene, and/or olivine) and have been proposed to be

108 diagnostic of the tectonic setting of serpentinization. This has unlocked opportunities to discern

109 the tectonic setting of mantle material exhumed in subduction complexes (Deschamps et al.,

110 2010, 2013; Peters et al., 2017).

111   Studies aiming to determine the tectonic source of formerly subducted serpentinites must

112   grapple with a nuanced, multi-staged history of serpentinization associated with pre-subduction

113   processes and the prograde and retrograde paths of subduction and exhumation (Figure 1).

114   Quantitatively determining the pressure (P), temperature (T), and time (t) conditions experienced

115   by exhumed serpentinites is challenging owing to high-variance phase assemblages, though

116   promising new approaches expand options for unlocking both temperature and time in these

117   rocks (Cooperdock & Stockli, 2016; Schwartz et al., 2020).

118   Complexities notwithstanding, there is a growing body of literature with examples of

119   provenance interpretations of serpentinites in exhumed subduction complexes, such as studies

120   assigning serpentinite associated with blueschist and eclogite to the downgoing plate (e.g.,

121   Cooperdock et al., 2018; Katzir et al., 2000; Li et al., 2004; Scambelluri et al., 1991; Shen et al.,

122   2015). Some studies have further resolved detail within the oceanic realm of the protolith,

123   interpreting exhumed serpentinites to derive from passive margin settings (e.g., Barnes et al.,

124   2014) or from abyssal transform faults (e.g., Cárdenas-Párraga et al., 2017). In many cases,

125   however, non-distinctive geochemical signatures, lack of relict phases, and heterogeneity of the

126   analyzed samples makes this level of resolution untenable. In other regions, the capturing of

127   hangingwall mantle via subduction erosion and other mass transfer mechanisms has been

128   identified (e.g., Bhat et al., 2019; Guice et al., 2021; Hattori et al., 2010; Lazar et al., 2021; Li et

129   al., 2018; Tewksbury-Christle et al., 2021; Wu et al., 2018). In yet other localities, serpentinites

130   from both the downgoing slab and the overlying mantle wedge are interpreted to be present,

131   implying complex slab-mantle interactions at a range of depths (e.g., Barnes et al., 2013; Blanco-

132   Quintero et al., 2011). These results generate discussion of deep tectonic slicing of slabs at depth,

133   degrees of mechanical mixing and styles of deformation (e.g., coherent nappe stacking, block-

134   and-matrix shear zones), and the processes of underplating and interaction with the mantle

135   material of the overriding plate.

136   In this contribution, we discern the tectonic origin and alteration histories of serpentinites

137   entrained in an Eocene high-pressure / low-temperature (HP/LT) subduction complex on the

138   island of New Caledonia (SW Pacific) and explore implications for the subduction and

139   exhumation history of the complex. The island of New Caledonia exposes an extraordinarily

140   complete Eocene subduction-obduction complex, including a high-temperature metamorphic

141   sole representing incipient stages of subduction initiation, an HP/LT subduction complex, and an

142 obducted supra-subduction zone ophiolite sequence (Maurizot et al., 2020a). These entities are

143 well-studied and world-class sites: the ophiolite comprises one of the world's largest continuous

144 mantle exposures and has provided invaluable insights into upper mantle processes and melt

145 transfer in the lower crust (e.g., Cluzel et al., 2016; Dupuy et al., 1981; Marchesi et al., 2009;

146 Pirard et al., 2013; Secchiari et al., 2016, 2018; Ulrich et al., 2010). The subduction complex

147 exposes slivers of oceanic and thinned continental lithosphere that experienced blueschist and

148 eclogite-facies metamorphism and has served as a location for investigation of subduction zone

149 fluid-rock interaction, volatile recycling, and slab-mantle interactions (e.g., Cluzel, 2021;

150 Spandler et al., 2008; Taetz et al., 2016, 2018). Despite the geodynamic importance of mantle

151 rocks throughout the subduction-obduction complex, studies addressing the distribution,

152 structure, texture, or geochemistry of meta-ultramafic rocks in the HP/LT complex are few

153 (Cluzel, 2021; Fitzherbert et al., 2004; Rawling & Lister, 2002; Spandler et al., 2008). These

154 studies arrive at differing conclusions regarding the source of the ultramafic material entrained in

155 the complex. To evaluate these differences and determine the tectonic provenance and

156 petrogenesis of ultramafic rocks in this subduction complex, our approach integrates field

157 observations, petrographic characterization, whole rock major and trace element geochemistry,

158 and stable isotope (O and H) geochemistry. Our results reveal the existence of two distinct

159 groups of serpentinites, which we interpret to indicate differing tectonic settings for their

160 protoliths.

161 **2 Geologic History of New Caledonia**

162   Located in the SW Pacific Ocean, the main island of New Caledonia exposes a dense

163 array of geologic terranes that record Paleozoic and Mesozoic histories through to present day: a

164 material archive that stands in contrast to many other islands in the Pacific, which are mostly

165 entirely Cenozoic in age and volcanic in origin (Cluzel et al., 2012; Maurizot et al., 2020a; Paris,

166 1981). The island (~16,000 km^2) is a rare emergent portion of the NW-SE trending submarine

167 Norfolk Ridge, a continental sliver located ~200-400 km southwest of active subduction at the

168 New Hebrides trench and comprising part of the largely submerged Zealandia microcontinent

169 (Figure 2a) (Crawford et al., 2003; Dubois et al., 1974; Lafoy et al., 2005; Mortimer et al., 2017).

170   New Caledonia exposes Late Carboniferous to Early Cretaceous basement rocks deriving

171 from the last accretionary stages of the active southern Gondwana margin (Figure 2b,

172   'undifferentiated "basement" terranes') (Aitchison & Meffre, 1992; Campbell et al., 1985;

173   Maurizot et al., 2020b); unless otherwise noted, we adopt the naming conventions of Maurizot et

174   al. (2020c) for New Caledonian rock units. In the Late Cretaceous, regional tectonic stresses

175   shifted, prompting divergence and rifting associated with the breakup of the Gondwanan

176   supercontintent. In the SW Pacific, the Tasman Sea and additional small ocean basins opened,

177   rifting ribbons of continental crust away from the eastern margin of Gondwana (Figure 2a) (e.g.,

178   the submarine Dampier Ridge and Lord Howe Rise, for instance; e.g., Bache et al., 2014; Davies

179   & Smith, 1971; Mortimer et al., 2018).

180   Plate convergence in the SW Pacific began in the Late Paleocene, with an intra-oceanic

181   NE-dipping subduction zone initiating at ~56 Ma, as recorded by the recrystallization age of

182   high-temperature amphibolite in the basal sole of the ophiolite, and further corroborated by

183   boninite and adakite series dikes that range from 55-50 Ma (Cluzel et al., 2006, 2012). Two

184   distinct packages of differing protolith types and P-T-t paths are recognized to have been

185   subducted and exhumed. The Pouébo Terrane (Figure 2c, d), which subducted first, consists of

186   fragments of oceanic lithosphere, and was followed by the Diahot-Panié Complex (Figure 2c, d),

187   which consists dominantly of metasediments and metavolcanic rocks (Cluzel et al., 1994; Clarke

188   et al., 1997; Maurizot et al., 2020b). The Pouébo Terrane reached peak eclogite facies

189   metamorphic conditions of ~2.2-2.4 GPa and ~550-600°C, equating to ~70-80 km burial depth,

190   by ~44 Myr (Pirard & Spandler, 2017; Spandler et al., 2005; Vitale-Brovarone et al., 2013). The

191   Diahot-Panié Complex ranges in grade from lawsonite-blueschist facies (~0.5 GPa, ~250°C) to

192   eclogite facies (~1.8-2.2 GPa, ~500-550°C) and likely reached peak conditions at ~38 Ma

193   (Cluzel et al., 2010; Pirard and Spandler, 2017; Potel et al., 2006; Vitale-Brovarone et al., 2018).

194   Exhumation is interpreted to have occurred in two stages: the first stage brought the Pouébo

195   Terrane rocks up to ~40-50 km, where they were juxtaposed with the Diahot-Panié Complex. In

196   the second stage, the two terranes exhumed together between 38-36 Myr, with the last gasps of

197   rapid exhumation taking place at ~34 Ma (Baldwin et al., 2007; Vitale-Brovarone et al., 2018).

198   The arrival and partial subduction of the buoyant continental-affinity Diahot-Panié Complex

199   effectively halted subduction and triggered the obduction of a large fragment of overlying fore-

200   arc mantle and oceanic crust from the modern-day Loyalty Basin onto Grande Terre (referred to

201   as the Peridotite Nappe and Poya Terrane, respectively; Brothers, 1974; Coleman, 1967).

202   2.1 Ultramafic rocks in New Caledonia

203   Ultramafic rocks occur in several discrete tectonic terranes on the island. We briefly

204 summarize the state of knowledge with respect to the provenance of these mantle rocks,

205 organized into tectonic groups.

206   2.1.1. The Peridotite Nappe: Massif du Sud and isolated ophiolitic klippen

207   The southern end of New Caledonia is dominated by the Massif du Sud, one of the

208 largest coherent exposures of mantle in the world (Figure 2b). This body is notably free of

209 significant tectonic overprint and remains attached to non-obducted lithospheric mantle in some

210 areas (Collot et al., 1987; Patriat et al., 2018; Prinzhofer & Nicolas, 1980). Additional remnants

211 of the large overthrust ophiolite dot the west coast of the island as small klippen. Most of the

212 mantle portion of the Nappe consists of harzburgite and dunite, with lherzolite being found in the

213 north (Belep, Poum, and Tiébaghi massifs; Moutte, 1982; Prinzhofer, 1981; Sécher, 1981; Ulrich

214 et al., 2010). The transition to the crustal sequence (pyroxenite, wehrlite, and gabbro) is

215 preserved on the southern portion of the island; the ophiolite lacks an uppermost dike complex

216 and pillow basalts (Maurizot et al., 2020c). The ophiolite is highly geochemically depleted

217 (Marchesi et al., 2009; Prinzhofer & Allègre, 1985), implying high degrees of melt production,

218 and this has made this body, and in particular, the continuous exposures in the Massif du Sud, a

219 globally-important site for studying melt transfer within the mantle and into the lower crust

220 (Marchesi et al., 2009; Pirard et al., 2013; Pirard & Hermann, 2015; Ulrich et al., 2010).

221   2.1.2. "Basement" ultramafic bodies of Gondwanan affinity

222   Serpentinites are reported in several Gondwanan basement terranes on the island.

223 Portions of these units experienced local low-grade (greenschist and/or lawsonite blueschist

224 facies) overprinting by Eocene HP/LT metamorphism. In some cases, serpentinites occur

225 amongst swaths of schist and contain tectonically-entrained oceanic crustal components (e.g.,

226 metamorphosed pyroxenite, gabbro, basalt, chert). This ultramafic material is proposed to derive

227 from an incoming oceanic plate that reached maximum depths of ~30-35 km beneath the

228 overriding Gondwanan plate ('Boghen Terrane'; Black, 1993; Cluzel & Meffre, 2002; Guérangé

229 et al., 1977; Maurizot et al., 2020b). In other localities within the basement, serpentinites appear

230 at faulted boundaries between coherent volcanic and abyssal sedimentary units (e.g., chert,

231 siltstones) (Maurizot et al., 2020b; Meffre et al., 1996). Published geochemical data on these

232 rocks are lacking and a protolith origin for these serpentinites is not known.

233   2.1.3. Foreland accretionary units

234   Ultramafic rocks are reported to cross-cut the Montagnes Blanches Nappe, a remnant

235 foreland fold-thrust belt preserved immediately west of the HP/LT complex (Koumac Terrane of

236 Cluzel et al., 1994). Early interpretations link these rocks to the underlying oceanic lithosphere

237 of the folded abyssal Montagnes Blanche Nappe sedimentary units (Brothers, 1974; Maurizot et

238 al., 1989). These rocks have alternatively been interpreted as overthrust remnants of the

239 Peridotite Nappe (Cluzel et al., 1995; Gautier et al., 2016; Maurizot, 2011), though a recent

240 systematic field description of serpentinite occurences in these units by Cluzel (2021) has

241 pointed out inconsistencies with this hypothesis, instead positing that they are part of the upper

242 plate. We emphasize that these hypotheses are based on field context, and published geochemical

243 studies of the serpentinites are lacking.

244   2.1.4. Ultramafic and hybrid "blackwall" rocks of the HP/LT complex

245   Serpentinites occur throughout the HP/LT metamorphic belt in the NE portion of the

246 island (Figure 2b-d). The exposed subduction complex is sprawling, spanning ~ 200 km long and

247 ~20 km wide. The far northern area of the complex is most often referred to as the Pam

248 Peninsula and exposes some of the best-preserved blueschists and eclogites in the complex

249 (Figure 2c). Though observations and descriptions of serpentinite outcrops exist (e.g., Black &

250 Brothers, 1977; Brothers & Blake, 1973; Cluzel, 2021; Lillie, 1975; Maurizot et al., 1989),

251 geochemical and textural studies of these ultramafic rocks are sparse (Fitzherbert et al., 2004;

252 Spandler et al., 2008), and the tectonic origin of this ultramafic material is debated. Spandler et

253 al. (2008) interpret a seafloor origin based on major and trace element geochemistry, stable

254 isotope measurements, and calculated temperatures of serpentinization for four serpentinite

255 samples in the far NE portion of the complex (Figure 2c). In contrast, Fitzherbert et al. (2004)

256 note the similarity of P-T paths and the existence of serpentinites across mapped terrane

257 boundaries (the Diahot Terrane and Pouébo Terrane of Cluzel et al. (1995) and Fitzherbert et al.

258 (2003)), interpreting this distribution as inconsistent with a downgoing slab origin, and instead

259 indicating interaction with the overlying mantle. This idea can be traced to earlier interpretations

260 that the ultramafic rocks exposed in the metamorphic complex today were part of a larger, now

261 extensively eroded overthrust "serpentinite sheet" that was incorporated from the hangingwall,

262 based on their consistent existence at the structurally highest levels of the complex (Rawling &

263 Lister, 2002). At a small, weathered massif outcropping in the SE portion of the complex, near

264 the town of Yambé (Figure 2d), Fitzherbert et al. (2004) interpret the ultramafic rocks to derive

265 from hangingwall mantle that was incorporated with the downgoing plate, subducting with it to

266 P-T conditions past the stability of antigorite. Critically, a geochemical link between the

267 serpentinites in the HP/LT complex and the interpretation of an overriding plate or supra-

268 subduction mantle origin is lacking.

269 **3 Sample Localities and Description**

270 Twenty-three serpentinites, one chlorite schist, and one talc schist were collected in the

271 HP/LT complex for the purposes of discerning their tectonic origin and fluid histories (Table S1;

272 Figure 2c, d). For comparison, four serpentinites were collected from outside the HP/LT

273 complex: two from the Boghen Terrane basement unit, one from the serpentinite sole at the base

274 of the Peridotite Nappe, and one from Kalaa-Gomen, a small klippe of the ophiolite on the

275 northwest coast (Figure 2b). Representative thin section photomicrographs are presented in

276 Figures S1-S4.

277 3.1 Field Context

278 Serpentinites in the northernmost portion of the HP/LT complex - defined here as an area

279 encompassed by the Pam Peninsula and extending south along the eastern coastal road (RPN7) to

280 the town of Balade - outcrop both as highly sheared and deformed lenses and as weathered

281 boulders on hillslopes (Figure 2c). In highly sheared areas, the serpentinites act as a matrix for

282 rounded pods of metamorphosed mafic, sedimentary, and other ultramafic rocks. With the

283 exception of a small quarry south of Col d'Amos (NC19-14) and a locality in the foothills north

284 of Ouégoa (NC18-26), all samples group spatially within the "Pouébo-Tiarì" unit of Vitale-

285 Brovarone et al. (2018). These samples are entirely encompassed within the "mélange

286 ophiolitique glaucophanite" unit of Maurizot et al. (1989) and fall variably within the "Pouébo

287 metabasite dominates" and "Diahot metabasite dominates" units of Fitzherbert et al. (2004).

288 Nine samples in the HP/LT complex are located SE of the northern domain samples. A

289 small (~ 0.5 km^2) ultramafic body crops out within blueschist and eclogite facies metamafic

290 rocks near the town of Yambé (Figure 2d). Prior studies reference this ultramafic body as

291 "Yambé" or "Yambé massif" (Fitzherbert et al., 2004; Spandler et al., 2008) and "Pwa Radèn"

292 (e.g., Cluzel, 2021). With permission from Kanak tribal leadership, we adopt the local name

293 given to this ultramafic body, "Poadja." In some areas of the Poadja Massif, relict magmatic

294 foliation is discernable at outcrop scale. Further south along the coast, two minor serpentinite

295 outcrops (within 1 km of each other) are exposed in shear zones in beach outcrops, juxtaposed

296 with meta-mafic blocks.

297   3.2 Petrographic Context

298   Serpentinites in the NE region are characterized by a dominant mineral assemblage of

299 antigorite + talc + magnetite (Table S1; Figures 3a, b and S1) and are > 90% serpentinized, with

300 rare relict chromian spinel (NC19-54), olivine (NC18-26B), and clinopyroxene (NC19-54).

301 Antigorite most commonly occurs as the dominant matrix phase, forming interlocking crystals

302 with variably preserved evidence for mesh texture. Some samples exhibit anhedral fine-grained

303 (<10 µm), recrystallized antigorite intergrown with talc and less commonly, tremolite (e.g.,

304 NC18-15C, NC18-15D, NC18-22A). In some cases, well-developed bastite pseudomorph

305 textures are evidenced by clumped aggregates of coarser-grained (~100-200 µm) euhedral

306 antigorite crystals accompanied by euhedral magnetite (~10's of µm) (e.g., NC19-86). Magnetite

307 additionally occurs as a matrix phase as single subhedral grains (~10's of µm up to ~1 mm) or as

308 polymineralic aggregates of smaller euhedral grains (~10's of µm). In deformed samples these

309 aggregates lie within the foliation (e.g., NC19-54, NC18-22A). In rare cases, relict chromite is

310 preserved in the core of grains mantled by ferritchromite and rimmed by magnetite (Figure 3a).

311 Two types of veins assemblages are present as cross-cutting networks: antigorite-magnetite

312 (NC19-86; Figures S1e, j) and antigorite-tremolite (NC19-94; Figure 3b).

313   Serpentinites in the SE region are characterized by the presence of relict phases

314 (orthopyroxene and olivine) and range between ~50 to ~90% serpentinized (Table S1; Figures

315 3c-f and S2). Samples from this locality notably lack talc (with the exception of NC19-158), and

316 four samples contain the Fe-Ni alloy awaruite as an accessory phase (Figure 3d). Oxide textures

317 show varying degrees of retrogression from skeletal chromite with magnetite rims (Figure 3c) to

318 chromian magnetite cores complexly replaced by magnetite (Figure 3e). Texturally-late brucite

319 veins are observed in one sample (NC18-39A). Mesh textures are less developed compared to

320 the complete pseudomorphing observed in the NE (Figure 3f).

321 **4 Methods**

322 4.1 Whole-rock major and trace element geochemistry (XRF, ICP-MS)

323 Twenty-three serpentinites, one talc schist, and one chlorite schist were analyzed for

324 whole-rock major and trace element compositions. Major element analyses were acquired by

325 lithium tetraborate fusion X-ray fluorescence methods at the GeoAnalytical Laboratory

326 (Washington State University, USA; WSU) and Franklin & Marshall College (Pennsylvania,

327 USA; F&M). Hand-picked fresh chips of each sample were powdered and fused on-site.

328 Additional details on preparation and fusion are reported in supporting information S1. At WSU,

329 major elements were analyzed on a ThermoARL Advant'XP+sequential X-ray fluorescence

330 (XRF) spectrometer using the preparation and analytical procedure described in detail by

331 Johnson et al. (1999) and Kelly (2018). At F&M, major elements were analyzed on a Malvern

332 PANalytical Zetium X-ray fluorescence spectrometer. Trace element concentrations were

333 acquired at WSU and the University of Rhode Island (USA). Digestions and dilution were

334 performed on-site. At WSU, trace elements were analyzed on an Agilent quadrupole ICP-MS

335 using the method of Knaack et al. (1994). Analytical precision is <5% for REE's and <10% for

336 the remaining elements. At University of Rhode Island, trace elements were analyzed on a

337 Thermo X-Series 2 quadrupole ICP-MS following the method of Savov et al. (2005).

338 Interference corrections were applied for TiO on Zn, Ba++ on Ga, and CrO on Nb. Analytical

339 precision is <5% for most elements, and precision for Cr and Ni was <2%. Two samples (NC18-

340 15D, NC18-49) were analyzed by all three labs (Table S2, S3).

341 4.2 Oxygen and hydrogen stable isotope geochemistry

342 Oxygen and hydrogen isotope compositions of serpentine, talc, and magnetite mineral

343 separates were measured at the University of Texas at Austin using a ThermoElectron MAT 253

344 mass spectrometer. Serpentinite samples were coarsely crushed and handpicked under a

345 binocular microscope to ensure purity of mineral separates. Handpicked serpentine and talc

346 grains were washed with dilute HCl to remove any trace carbonate material. Approximately 2.0

347 mg of each mineral separate were analyzed using the laser fluorination method of Sharp (1990)

348 in which samples were heated by a CO_2 laser in the presence of a BrF_5 atmosphere to liberate

349 oxygen. Liberated oxygen was cryogenically purified and analyzed as O_2. Precision and accuracy

350 of oxygen analyses were verified through garnet standard UWG-2 ($\delta^{18}O$ = +5.8‰) (Valley et al.,

351 1995), in-house olivine standard San Carlos ($\delta^{18}O$= +5.2‰), and in-house quartz standard

352 Lausanne-1 ($\delta^{18}O$= +18.1‰). All $\delta^{18}O$ values are reported relative to SMOW, where the $\delta^{18}O$

353 value of NBS-28 is +9.7‰. The error on each $\delta^{18}O$ analysis is ±0.1‰, based on the long-term

354 average of standard analyses. Seven samples were run in duplicate to ensure reproducibility and

355 consistency over four analytical runs.

356   Hydrogen isotope ratios were measured using the method of Sharp et al. (2001). Samples

357 were hand-powdered using an agate mortar and pestle. Approximately 1 mg of each sample were

358 loaded into silver foil capsules, dried under vacuum at 70° C for 24 h, transferred to a Costech

359 zero-blank autosampler, and flushed with He gas. Samples were measured by continuous-flow

360 mass spectrometry using a ThermoElectron TC/EA (high-temperature conversion elemental

361 analyzer) coupled to the ThermoElectron MAT 253 mass spectrometer. Four internationally

362 referenced and certified standard materials (IAEA-CH7, NBS-22, USGS-57, USGS-58) and one

363 in-house working glass standard were analyzed with the samples throughout the run. Raw δD

364 values were corrected for instrumental drift and normalized to SMOW using a calibration curve

365 generated from the measurements of the standard reference materials. Error based on the

366 reproducibility of standards measured in the analytical runs is ± 2‰. δD values referenced in the

367 text and plotted are the mean of two individual replicate analyses.

368   4.3 Raman spectroscopy

369   Raman spectra were acquired on eight serpentinite samples (two samples from Group I,

370 four samples from Group II, one from the Boghen Terrane, and the sample from Kalaa-Gomen).

371 In situ spot analyses on 30 µm polished thin sections were acquired on a Witec Alpha 300R

372 confocal Raman microscope at the Characterization Facility, University of Minnesota. The

373 confocal Raman microscope is equipped with a UHTS300 spectrometer and DV401 CCD

374 detector. Spectra were acquired with a frequency doubled Nd:YAG 532 nm laser, a 1,800 g/mm

375 grating, and a 100x objective. Spot size was ~1 µm in diameter. Each spectra resulted from the

376 average of two 20 s acquisitions to optimize the signal/noise ratio. Two spectral intervals were

377 measured: a low-wavenumber region (100-1200 cm^{-1}) for structural bonding characterization and

378   a high-wavenumber region (3300- 4050 cm^{-1}) for characterization of hydroxyl groups. A range

379   of relevant textural settings were targeted in each thin section (e.g., serpentine matrix, veins, and

380   pseudomorphs). Data were processed using the WITec Project Five+ software. Raw spectra

381   underwent background subtraction followed by smoothing using a third-order polynomial

382   Savitzsky-Golay filter. Serpentine species were identified by comparison to previously published

383   data (Auzende et al., 2004; Groppo et al., 2006, Petriglieri et al., 2015; Tarling et al. 2018).

5 Results

385   5.1 Whole rock major elements

386   Serpentinites from the HP/LT complex display differences in SiO_2, Al_2O_3, MgO, and

387   CaO contents and cluster in two distinct geochemical groupings (Table 1; Figure 4). One group

388   of serpentinites (n = 15; herein referred to as Group I) is depleted in MgO compared to other

389   serpentinites (herein referred to as Group II, n = 9): 34.05 ± 3.34 wt% in Group I versus 40.19 ±

390   3.20 wt% in Group II. Group I serpentinites contain higher Al_2O_3 contents than Group II (2.34 ±

391   1.47 versus 0.75 ± 0.34 average wt%, respectively). Group I samples are generally elevated in

392   SiO_2 compared to Group II, though they overlap within uncertainty, with averages of 44.96 ±

393   5.23 wt% and 41.13 ± 2.13 wt%, respectively. $Fe_2O_{3(T)}$ contents do not vary significantly

394   between the two groups, though samples in Group I display a broader range of variation (7.92 ±

395   1.85 wt % versus 7.38 ± 0.68 wt% for Group II). CaO concentrations in the HP/LT serpentinites

396   overall are <0.5 wt%, with the exception of the three least serpentinized samples (Table 1 and

397   Figure S2); all Group II samples located in the SE portion of the complex (0.53, 0.57, and 1.08

398   wt%). The ophiolitic reference sample from Kalaa-Gomen (NC19-178) has 0.66 wt% CaO.

399   Samples have a range of loss-on-ignition (LOI) values, from 5.25 to 12.64%, indicating varying

400   degrees of hydration, with an overall average of 10.43% ± 2.77%. Na_2O, K_2O, and P_2O_5 are

401   close to or below detection limits.

402   The chlorite schist (NC19-85) is elevated with respect to the talc schist (NC19-169) in

403   Al_2O_3, $Fe_2O_{3(T)}$, and TiO_2, whereas the talc schist is elevated in SiO_2 and MgO (Table 1). The

404   chlorite schist (NC19-85) has CaO and P_2O_5 contents of 2.47 and 1.88 wt%, respectively, due to

405   the presence of ~5% modal abundance apatite.

406   5.2 Whole-rock trace elements

407   Group I and II samples are overall depleted in trace elements with respect to primitive

408   mantle (Figure 5a, b). Both groups show enrichments in Cs and Pb and negative Rb and Nb

409   anomalies. Group I samples show marked Th and U enrichments and a strong negative Sr

410   anomaly compared to Group II. Group I samples show greater overall enrichment enrichment in

411   middle to heavy rare-earth elements (M-HREE; Sm, Eu, Gd, Tb, Dy, Y, Ho, Er, Yb, Lu)

412   compared to Group II (Figure 5c-f). Differences in the pattern of light-rare earth elements

413   (LREE; La, Ce, Pr, Nd, Pm) to HREE most clearly distinguish the two groups ($La_N/Ho_N = 1.66 \pm$

414   1.33 for Group I and 6.59 ± 1.92 for Group II; normalized to primitive mantle).

415   The chlorite schist (NC19-85) and talc schist (NC19-169) samples display distinct

416   enrichments in trace elements. The chlorite schist is enriched in Li, Sc, V, Cu, Zn, Sr, Y, Zr, Nb,

417   Ba, La, Ce, Pr, Nd, Sm, Eu, Tb, Gd, and Dy relative to the talc schist. Conversely, the talc schist

418   is enriched in Cr and Ni relative to the chlorite schist.

419   5.3 Stable isotope geochemistry

420   5.3.1 Oxygen isotopes

421   Serpentine from the high-pressure complex displays a wide range of oxygen isotope

422   values: from +5.1 to +10.2‰ (Table 2 and Figure 6a, b). Group I samples are between +6.7 and

423   +10.2‰, whereas Group II samples are between +5.1 to +8.0‰. The mean serpentine $\delta^{18}O$ value

424   for Group I is $+8.7 \pm 0.9$‰ (n = 15) and $+5.5 \pm 0.4$‰ (n = 9) for Group II. A Group II sample

425   (NC18-39A) shows a 1‰ increase in $\delta^{18}O$ between serpentine extracted from the sample interior

426   (+5.4‰) and serpentine from an outer altered rind (+6.4‰). $\delta^{18}O$ sample replicates reproduced

427   within 0.2‰, with the exception of two samples, NC18-15D and NC18-22A, which show

428   variability of 1.8‰ and 0.9‰, respectively. Comparative samples from the Boghen Terrane,

429   Kalaa-Gomen Massif, and the Peridotite Nappe serpentinite sole have $\delta^{18}O$ values of +7.5‰

430   (average of two samples), +5.7‰, and +6.5‰, respectively.

431   Oxygen isotopes in magnetite were measured in three samples from the NE part of the

432   complex: a serpentinite (NC19-86), a chlorite schist (NC19-85), and a talc schist (NC19-169).

433   The $\delta^{18}O$ value of magnetite is +1.5‰ in the serpentinite and +2.4‰ in the chlorite schist. In the

434   talc schist, two separate grain size fractions were measured, 125-250 µm and >710 µm. The 125-

435 250 µm grain size fraction has a $\delta^{18}O$ value of +4.3‰, whereas the >710 µm has a value of +4.6

436 µm. Oxygen isotopes in talc from two samples (one from the SE and one from the NE), were

437 +9.0 and +10.5‰, respectively.

438   5.3.2. Hydrogen isotopes

439   δD values of serpentine in the high-pressure complex range from −79 to −29‰ (Figure

440 6a). Group I sample range between -76 and -29‰ and Group II samples range between -79 and -

441 39‰. The mean serpentine δD value is -44 ± 10‰ (n = 30) for Group I and -52 ± 14‰ (n = 18)

442 for Group II. The relatively large standard deviation in both groups derives from three samples

443 (one in Group I and two in Group II) that have δD values between -79 and -70‰. Without these

444 values, the remaining samples from Group I exhibit a more restricted range of δD values from -

445 49 to -32, with an average of -41± 5‰ (n = 28). Group II ranges from -59 to -39‰, with an

446 average of -46 ± 8‰ (n = 14). Comparative samples from the Boghen Terrane, Kalaa-Gomen

447 Massif, and the Peridotite Nappe serpentinite sole have δD values of -81‰ (average of two

448 samples), -85‰, and -82‰, respectively. Across all samples, δD sample replicates reproduced

449 within 2‰, on average.

450   5.4 Raman spectroscopy

451   Serpentine has three primary structural polymorphs that vary over *P-T* space: the low-T

452 form is chrysotile, followed by lizardite and antigorite. Diagnosing the polymorph in exhumed

453 serpentinites can aid in retrieving information about its prograde and/or retrograde path and

454 alteration. Serpentine in two samples from Group I and four samples from Group II was found to

455 be antigorite (only) in contrast to lizardite identified in comparative samples from the Boghen

456 Terrane basement unit and Kalaa-Gomen massif in the Peridotite Nappe (Figure 7 and S5-6).

457   5.5 Categorization of New Caledonia HP/LT serpentinites

458   As presented above, serpentinites in the high-pressure metamorphic complex cluster in

459 two distinct groups, revealed by petrography, major and trace element, and stable isotope

460 geochemistry. The significance of these groups and their composition will be discussed in the

461 next section. The results are summarized as follows:

462   Group I: >90% serpentinized, common assemblage of antigorite + talc + magnetite,

463   lower, more homogeneous MgO contents (relative to Group II), $\delta^{18}O$ values between

464   +6.7‰ and 10.2‰, flat M-HREE patterns;

465   Group II: ~50-90% serpentinized, notable preservation of orthopyroxene and olivine and

466   distinct presence of awaruite, higher MgO contents (relative to Group I), $\delta^{18}O$ between

467   +5.1‰ and +8.0‰, distinct curved LREE to HREE pattern compared to Group I.

468 **6 Discussion**

469   Prior studies of serpentinites in the HP/LT terrane of New Caledonia have disagreed on

470 the tectonic origin of the mantle protolith (Fitzherbert et al., 2004; Spandler et al., 2008). Our

471 results indicate the existence of two geochemically-distinct groups of serpentinites in the

472 blueschist- to eclogite-facies portions of the complex. This heterogeneity may derive from (1)

473 geochemical differences in the original mantle source material, (2) processes occurring

474 throughout serpentinization, (3) subsequent metamorphism during subduction and exhumation,

475 or (4) a combination of the above.

476   6.1 Assessing major and trace element mobility

477   The major and trace element compositions of exhumed serpentinites result from (a) the

478 original composition and melt history of the parent peridotite, (b) fluid-rock reactions occurring

479 at the source location for serpentinization (e.g., seafloor, mantle wedge), and (c) fluid-rock

480 reactions occurring during subduction and exhumation. Extensive study of variably serpentinized

481 abyssal and mantle wedge peridotites and serpentinites in exhumed subduction complexes has

482 demonstrated high-field strength elements (HFSE; Nb, Ta, Zr, Hf) and REE are generally (but

483 not wholly) immobile during serpentinization and aqueous fluid alteration and can be used to

484 discern information about protolith composition and pre-serpentinization magmatic processes

485 (e.g., Deschamps et al., 2013; Kodolányi et al., 2012; Niu, 2004; Parkinson & Pearce, 1998;

486 Savov et al., 2005; Scambelluri et al., 2004). By contrast, major elements and FME (e.g., Li, Pb,

487 U, Cs, Sr, Ba) must be interpreted with additional caution, as serpentinization is not isochemical

488 with respect to some of these elements and/or they are more readily mobilized during fluid

489 interaction over the course of subduction and exhumation (e.g., Cannao et al., 2016; Malvoisin,

490 2015; Peters et al., 2017). With this in mind, we first assess the degree to which major and trace

491   elements in the New Caledonia HP/LT serpentinites may have been mobilized, or whether they

492   otherwise retain faithful geochemical signatures of their initial mantle protolith.

493   Ratios of immobile trace elements (e.g., LREE, HFSE vs HREE; La/Yb, Zr/Yb, Hf/Yb,

494   Nd/Yb) are observed to vary independently of LOI for both groups of HP/LT serpentinites

495   (Figure 8). This result, combined with the observed similarity of intra-group REE patterns for

496   Groups I and II (Figure 5), provides initial confidence in the retention of protolith mantle

497   geochemical signatures for these elements (c.f. Savov et al., 2005). For further assessment, plots

498   of LREE (La, Ce, Pr, Nd) versus HFSE (Nb, Ta, Zr, Hf) elucidate whether a common process or

499   processes were responsible for the enrichment or depletion of these groups of elements (c.f. Niu,

500   2004). It is worth noting that LREE are more readily mobilized by aqueous fluids than MREE

501   and HREE, so this test is a conservative indicator of trace element and REE immobility.

502   Group I LREE's are almost entirely decoupled from HFSE (Figure 9). Indications of a

503   once-coupled relationship between the LREE and HFSE are partially retained in a few samples,

504   as indicated by consistent low abundances that correlate somewhat more linearly (but to a degree

505   no longer statistically significant). More often, however, these samples have been overprinted by

506   a process or processes that added LREE (Figure 9). In contrast, statistically significant

507   correlations are observed for Group II serpentinites between the LREE and Zr and Nb ($R_{LREE-Zr}$

508    $_{(avg)} = 0.942 \pm 0.027$, $R_{LREE-Nb (avg)} = 0.678 \pm 0.024$) (Figure 9). Notably, LREE appear decoupled

509   from Ta and Hf ($R_{LREE-Ta (avg)} = -0.267 \pm 0.037$, $R_{LREE-Hf (avg)} = 0.308 \pm 0.106$). Of the HFSE, Zr

510   and Nb (here coupled with LREE) are lighter in mass than Ta and Hf. Niu (2004) addresses the

511   potential importance of mass-dependent effects on the observed magmatic fractionation of Nb/Ta

512   and Zr/Hf ratios in global abyssal peridotite datasets, proposing that elevated LREE abundances

513   could be produced or enhanced by mass-dependent transfer rates, but notes that observational

514   tests of this hypothesis are necessary. These data possibly corroborate this hypothesis,

515   particularly given that none of the four elements correlate with LOI. However, the possibility

516   that this decoupling reflect later aqueous alteration processes cannot be ruled out, though it is

517   unclear whether subduction zone temperatures would be sufficiently high enough to mobilize

518   these HFSE.

519   In summary, both groups show variations in trace and REE independent of LOI - a

520   valuable first indicator for retained magmatic trace element signatures. Strong correlation

521   between LREE and HFSE in Group II samples bolsters this argument. Observed mobility in

522   Group I LREE does not preclude us from utilizing the REE data to ascertain a tectonic setting for

523   the mantle protolith, as there is no demonstrated correlation between the M-HREE with LOI;

524   though care must be taken in interpreting the significance of LREE abundances and their

525   contribution to the overall shape of L-HREE trends.

526   6.2 Serpentinization processes and fluid interactions

527   Within the ultramafic rock system, oxygen and hydrogen isotopes have been widely

528   utilized to discern the source of serpentinizing fluids and subsequent post-serpentinization fluid

529   histories (e.g., Alt and Shanks, 2006; Burkhard and O'Neil, 1988; Früh-Green et al., 1990, 2001;

530   Kyser et al., 1999). Ultimately, stable isotope values are the amalgamated product of several

531   important variables – the stable isotope composition of the interacting fluid, the fluid-rock ratio,

532   the composition of the reactive solid phase(s) in the system, and the fractionation factors

533   between these components and the fluid. Stable isotope data for the HP/LT serpentinites indicate

534   differing serpentinizing fluid compositions and/or T conditions for serpentinization between the

535   two groups of samples. Group II $\delta^{18}O$ values (+5.5 ± 0.4‰) cluster tightly around the typical

536   upper mantle value (+5.4‰), indicating little to no additional alteration of the primary mantle

537   signature during serpentinization, subduction, or exhumation (Figure 6). This assessment is

538   supported by recent stable isotope modeling of the ophiolite that indicates limited interaction of

539   slab sediment-derived fluids within New Caledonia's mantle wedge during the initial phases of

540   subduction (Ulrich et al., 2020). Moreover, these authors discern, on the basis of Sr isotopes and

541   low abundances of sediment-derived elements such as As and Sb (Deschamps et al., 2013), that

542   the input of subducted sediments was low at the time of serpentinization of the mantle wedge,

543   and instead the dehydration of altered oceanic crust comprised the dominant fluid source.

544   Though our data do not include measurements of As and Sb, we observe similarly low degrees of

545   elements typically regarded as classic markers of serpentinization in the mantle wedge via

546   shallow dewatering fluids released from slab sediments and lower-T metamorphic dehydration

547   reactions (e.g., Cs, Sr, Rb, Li) (Figure 1). The narrow range of $\delta^{18}O$ values across all degrees of

548   hydration (as indicated by LOI %) supports a consistent temperature of serpentinization and/or a

549   consistent fluid composition (Figure 6). The close relation between the $\delta^{18}O$ values of Group II

550 serpentinites and antigorite within slab-derived tremolite-antigorite veins in the Peridotite Nappe
551 also supports a similar source for serpentinizing fluids.

552 By contrast, Group I samples show increased and variable $\delta^{18}O$, which may be
553 interpreted to reflect: (1) interaction with metamorphic fluids with a sedimentary component, (2)
554 differing temperatures of serpentinization in a seafloor environment, or (3) a mix of these two
555 influences. Both end-member explanations are plausible - all but two Group I samples are
556 located within 2 km of interpreted boundaries between the Pouébo Terrane and the sediment-
557 dominated Diahot-Panié Complex. Moreover, significant fluid fluxing is evident for these
558 samples given major element concentrations, which show a clear deviation from the terrestrial
559 array as a function of increasing modal abundances of talc (Figure 6). The development of talc-
560 rich serpentinite and hybrid rock assemblages in the NE portion of this HP/LT complex has been
561 attributed to high fluid flux-driven metasomatism at depth and is the source of a separate study
562 by Spandler et al. (2008).

563 A suite of discriminant fluid-mobile trace element diagrams complements our
564 interpretation of significantly different fluid and tectonic histories for Groups I and II. FME's in
565 serpentinites are an sensitive recorder of fluid interactions, in part because peridotite protoliths
566 are depleted in these elements, so increases in their concentration are attributable to fluid
567 interactions (e.g., Peters et al., 2017; Scambelluri et al., 2019). Recent re-analysis and refinement
568 of compiled serpentinite data has also demonstrated that these elements are sensitive tracers of
569 the tectonic environments of serpentinization (Peters et al., 2017). For the New Caledonia HP/LT
570 serpentinites, plots of fluid-mobile element enrichments (e.g., Ba, Cs) relative to fluid-immobile
571 elements such as Yb show compelling trends (Figure 10). Ba in Groups I and II (Figure 10a)
572 indicates a mantle wedge origin for Group II samples, which are relatively enriched compared to
573 Group I samples, which fall squarely within the field for mid-ocean ridge serpentinites. This
574 trend is not observed for Cs, which we interpret as becoming enriched in Group I samples during
575 intense fluid flow and deformation, and possible interaction with neighboring sediments from the
576 Diahot-Panié Complex. Cs in Group II falls lower than compiled forearc data, commensurate
577 with our interpretations for limited influence of sedimentary-derived fluids based on stable
578 isotope data. This Cs depletion is apparent in a plot of Cs vs fluid-immobile Yb, though the data
579 in this context better overlap with the range of observed values for drilled or dredged forearc
580 samples, and the Group I samples correlate well with the compiled mid-ocean ridge dataset.

581   Trends between U and Th enrichments show that Group I samples lack the seafloor U

582   enrichment that defines some (but not all) of the mid-ocean ridge dataset (Figure 10c).

583   6.3 Tectonic origin of New Caledonia HP/LT-complex serpentinites

584   Given evidence presented above, we interpret differing tectonic origins for serpentinites

585   in the NE and SE portions of the terrane:

586   Group I: Found dominantly in the NE part of the complex, these serpentinites are

587   remnants of lithospheric mantle from the downgoing plate (Figure 11a). They experienced

588   progressive serpentinization on or near the ocean floor prior to subduction and then faced high-

589   fluid-rock ratios, metasomatism leading to the development of talc-bearing assemblages, and

590   deformation in the subduction channel. The formation of these rocks was possibly associated

591   with the development of several major shear zones, which facilitated slicing of the complex at

592   depth and have given parts of the Pouébo Terrane the appearance of a mélange. This geodynamic

593   interpretation is supported by recent reevaluations of the architecture of the belt (Maurizot et al.,

594   2020c; Vitale-Brovarone et al., 2018) and consistent with our observation of the proximity of

595   serpentinites to major unit boundaries and faults (e.g., NC18-22A, NC18-26B, NC19-63) and

596   their prevalence as matrix hosts for dismembered blocks in km-scale shear zones. It is plausible

597   that the pre-subduction abyssal environment for the future Pouébo Terrane ultramafic rocks was

598   characterized by lower-T conditions, consistent with stable isotope data presented here (Figure

599   6b). Seismic imaging of the New Caledonia Basin to the west of the island has revealed thinned

600   continental crust interpreted to consist of a significant volume of serpentinized upper mantle and

601   resembling the structure of crust formed during slow, low-T amagmatic seafloor spreading

602   (Klingelhoefer et al., 2007).

603   Group II: These samples cluster in the southern portion of the complex, comprising the

604   Poadja Massif. Incorporated onto the top of the slab prior to the onshore obduction of the

605   Peridotite Nappe, this slice of mantle wedge was subducted and exhumed intact, with striking

606   preservation of relict minerals and comparatively minor degrees of fluid alteration, as evidenced

607   by $\delta^{18}O$ isotope values that remain near the pristine upper mantle average and lower degrees of

608   LREE and FME (Cs, U, Pb, and Sr) enrichment relative to other global forearc mantle settings

609   (Figure 11b). We interpret these data to reflect the lack of a strong sedimentary-derived fluid flux

610   during serpentinization of these rocks in their mantle wedge setting, and this is supported by

611   independent evidence from analyses of slab-derived tremolite vein assemblages at the base of the

612   obducted Peridotite Nappe (Cluzel et al., 2020). This muted enrichment of LREE and FME also

613   indicates that later overprinting by meta-sedimentary derived fluids during transport within the

614   subduction channel was limited. The samples display characteristic REE patterns that remarkably

615   resemble the harzburgitic mantle of New Caledonia's Peridotite Nappe (Figure 11b).

616   6.4 Relevance for the architecture of the New Caledonia HP/LT complex

617   In the NE portion of the complex, our combined field observations and geochemical data

618   suggest that Group I (downgoing-slab) serpentinites: (a) overwhelmingly appear in the highest-

619   grade portions of the complex, within the oceanic affinity Pouébo Terrane, and (b) show distinct

620   geochemical markers for high degrees of fluid interaction and deformation during subduction

621   and/or exhumation (i.e., elevated $\delta^{18}O$ values, decoupled LREE's, enriched FME's such as Cs).

622   These observations are broadly consistent with recent re-interpretations of this subduction

623   complex by Vitale-Brovarone et al. (2018), which posits the existence of two accretionary

624   domains preserved in a nappe-type belt formed by progressive underplating and stacking. The SE

625   portion of the complex is far less studied: the existence of Group II mantle wedge serpentinites

626   carries implications for the structural level of the subduction zone sampled by rocks in the

627   immediate vicinity of the Poadja Massif. The existence of mantle wedge serpentinites adds to

628   independent observations made by Cluzel (2021) of the presence of hangingwall mafic rocks in

629   lower-grade portions of the Diahot-Panié Complex further north. Integrating these observations

630   into a unified subduction complex-scale picture of preserved lithostratigraphy requires further

631   studies. Nevertheless, these observations provide compounding evidence for greater structural

632   coherence within the New Caledonia subduction "mélange" than has previously been posited.

633   6.5 Implications for slab-mantle wedge interactions

634   Our finding of mantle wedge material within the exhumed HP/LT complex on New

635   Caledonia provides direct evidence for interaction with the hangingwall mantle in this Eocene

636   subduction zone. Geochemical modeling has demonstrated the importance of the mantle wedge

637   as a reservoir for water and fluid-mobile elements that contribute to arc magma isotope and

638   elemental budgets (e.g., Debret et al., 2019, 2020; Ribeiro & Lee, 2017; Savov et al., 2005,

639   2007), and there are strong conceptual bases, observational evidence, and modeled predictions

640   for the incorporation and downward dragging of mantle wedge material in some subduction

641   zones (e.g., Bebout & Barton, 2002; Bebout, 2007; Cloos & Shreve, 1988; Malatesta et al., 2012;

642   Peacock & Hyndman, 1999; Savov et al., 2005). New Caledonia provides an important natural

643   rock record comparison for geodynamic modeling: particulary for outcomes in which

644   exhumation of downgoing plate mantle *and* overlying mantle wedge material is predicted (e.g.,

645   Gerya & Stockert, 2006; Gerya et al., 2002; Malatesta et al., 2012). In these cases, the rheology

646   of serpentine and evolving hydration state of the ultramafic material are critical to the locus of

647   the weak, low-viscosity subduction channel and eventual migration of the plate interface

648   upwards into the mantle wedge. The development of talc in the overlying mantle assemblage

649   provides a possible physiochemical mechanism for this migration, and its stability in the

650   ultramafic chemical system may even control the depth of plate decoupling (Peacock & Wang,

651   2021).

652   This hypothesis is interesting to consider in light of our observations of relatively

653   undeformed, nearly talc-absent mantle wedge assemblages observed in the SE and deformed,

654   talc-bearing downgoing mantle assemblages in the NE. A local bulk assemblage-induced

655   rheological switch in the mantle wedge and consequent avulsion of the slab-mantle interface

656   could be broadly consistent with the results presented here. In this scenario, deformation rapidly

657   re-localizes in the new region of weakness and a stranded slice of mantle wedge material is

658   incorporated atop the downgoing slab. This mantle may feasibly have already experienced partial

659   high-T serpentinization from fluxed aqueous dehydration fluids (consistent with stable isotope

660   data presented here) and might largely escape prolonged intense deformation, because the

661   subduction interface was localized first below, and then above the mantle wedge slice in

662   rheologically weaker assemblages (consistent with observed petrographic textures and field

663   observations).

664   Tectonic erosion of the mantle wedge, and the type of HP/LT ultramafic material it

665   produces, contrasts markedly with the processes that retrieve lithospheric mantle under

666   subducted oceanic crust. Slicing and nappe development along lithologic (rheologic)

667   heterogeneities in the slab are commonly invoked as ways by which downgoing mantle may be

668   incorporated into the subduction channel (e.g., Angiboust & Agard, 2010; Guillot et al., 2009;

669   Hermann et al., 2000; Wakabayashi, 1992). During these processes, the mantle is juxtaposed and

670   reacts with sedimentary and mafic components of the slab, as reflected by stable isotope

671   compositions and Si and trace element enrichments. The data presented here for New

672   Caledonia's NE serpentinites reflect this type of metamorphic history, including their spatial

673   concentration within the oceanic Pouébo Terrane and proximity to inferred boundaries with the

674   Diahot-Panié Complex.

675   **7 Summary**

676   In this study, we resolve two petrographically, geochemically, and spatially distinct

677   groups of serpentinites within New Caledonia's HP/LT subduction complex. Our work

678   reconciles existing disparate interpretations of the tectonic origin of these rocks, with

679   implications for slab-mantle interaction during subduction. In the NE portion of the complex

680   serpentinite geochemistry and field relations indicate a downgoing slab origin. By contrast, the

681   chemistry of serpentinites in the SE points to a mantle wedge origin, indicating incorporation and

682   subduction of upper plate material. The exhumation of serpentinites from these two different

683   geodynamic settings within a single HP/LT complex is not common in the natural rock record,

684   but predicted for certain modeled scenarios as a function of serpentine rheology and hydration

685   state, among other subduction parameters. The mineralogy, chemistry, and spatial distribution of

686   New Caledonia's HP/LT serpentinites hold clues for understanding the architecture of this

687   subduction complex, and more broadly, for the modeled dynamics of deep tectonic slicing of

688   subducted oceanic plates and mass transfer between the upper and lower plates.

689   **8 Acknowledgments**

707   **9 Open Research**

708   Data Availability Statement

709   To ensure the accessibility and discoverability of the samples used in this research, and to

710   align with the National Science Foundation's guidelines of effective data practices, all samples

711   used in this study have been registered with IGSN Global Sample Numbers through the System

712   for Earth Sample Registration (SESAR). SESAR is maintained by the GeoInformatics Research

713   Group of the Lamont-Doherty Earth Observatory. https://www.geosamples.org/ Accessed 17

714   February 2022. Whole rock and stable isotope data presented in this paper will be submitted for

715   archival in EarthChem data repository prior to paper acceptance.

716   **10 References**

717   Aitchison, J., & Meffre, S. (1992). New Caledonia: a tectonic collage in the southwest Pacific.

718   In 29th International Geological Congress (Kyoto, 24 August-3 September 1992).

719   Alt, J., & Shanks III, W. (2006). Stable isotope compositions of serpentinite seamounts in the

720   Mariana forearc: Serpentinization processes, fluid sources and sulfur metasomatism. Earth and

721   Planetary Science Letters, 242(3-4), 272-285. https://doi.org/10.1016/j.epsl.2005.11.063

722   Alt, J. C., Schwarzenbach, E. M., Früh-Green, G. L., Shanks, W. C., Bernasconi, S. M., Garrido,

723   C. J., et al. (2013). The role of serpentinites in cycling of carbon and sulfur: Seafloor

724   serpentinization and subduction metamorphism. Lithos, 178, 40–54.

725   https://doi.org/10.1016/j.lithos.2012.12.006

726   Angiboust, S., & Agard, P. (2010). Initial water budget: The key to detaching large volumes of

727   eclogitized oceanic crust along the subduction channel? Lithos, 120(3–4), 453–474.

728   https://doi.org/10.1016/j.lithos.2010.09.007

729   Auzende, A.-L., Daniel, I., Reynard, B., Lemaire, C., & Guyot, F. (2004). High-pressure

730   behaviour of serpentine minerals: a Raman spectroscopic study. Physics and Chemistry of

731   Minerals, 31(5), 269-277. https://doi.org/10.1007/s00269-004-0384-0

732   Bache, F., Mortimer, N., Sutherland, R., Collot, J., Rouillard, P., Stagpoole, V., & Nicol, A.

733   (2014). Seismic stratigraphic record of transition from Mesozoic subduction to continental

734   breakup in the Zealandia sector of eastern Gondwana. Gondwana Research, 26(3-4), 1060-1078.

735   https://doi.org/10.1016/j.gr.2013.08.012

736   Baldwin, S. L., Rawling, T., & Fitzgerald, P. G. (2007). Thermochronology of the New

737   Caledonian high-pressure terrane: Implications for middle Tertiary plate boundary processes in

738   the southwest Pacific. Convergent Margin Terranes and Associated Regions: A Tribute to W.G.

739   Ernst, 419(06), 117-134. https://doi.org/10.1130/2006.2419(06)

740   Barnes, I., & O'Neil, J. R. (1969). The relationship between fluids in some fresh Alpine-type

741   ultramafics and possible modern serpentinization, Western United States. GSA Bulletin,

742   80(October), 1947–1960. https://doi.org/10.1130/0016-7606(1969)80[1947:TRBFIS]2.0.CO;2

743   Barnes, J. D., Beltrando, M., Lee, C.-T. A., Cisneros, M., Loewy, S., & Chin, E. (2014).

744   Geochemistry of Alpine serpentinites from rifting to subduction: A view across paleogeographic

745   domains and metamorphic grade. Chemical Geology, 389, 29-47.

746   https://doi.org/10.1016/j.chemgeo.2014.09.012

747   Barnes, J. D., Eldam, R., Lee, C.-T. A., Errico, J. C., Loewy, S., & Cisneros, M. (2013).

748   Petrogenesis of serpentinites from the Franciscan Complex, western California, USA. Lithos,

749   178, 143-157. https://doi.org/10.1016/j.lithos.2012.12.018

750   Barnes, J. D., Manning, C. E., Scambelluri, M., & Selverstone, J. (2018). The Behavior of

751   Halogens During Subduction-Zone Processes. https://doi.org/10.1007/978-3-319-61667-4_8

752   Bebout, G. E. (2007). Metamorphic chemical geodynamics of subduction zones. Earth and

753   Planetary Science Letters, 260(3–4), 373–393. https://doi.org/10.1016/j.epsl.2007.05.050

754   Bebout, G. E., & Barton, M. D. (2002). Tectonic and metasomatic mixing in a high-T,

755   subduction-zone mélange—insights into the geochemical evolution of the slab–mantle interface.

756   Chemical Geology, 187(1–2), 79–106. https://doi.org/10.1016/S0009-2541(02)00019-0

757   Behr, W. M., & Bürgmann, R. (2021). What's down there? The structures, materials and

758   environment of deep-seated slow slip and tremor. Philosophical Transactions of the Royal

759   Society A: Mathematical, Physical and Engineering Sciences, 379(2193), 20200218. https://

760   doi.org/10.1098/rsta.2020.0218

761   Bhat, I. M., Ahmad, T., & Subba Rao, D. V. (2019). Origin and evolution of Suru Valley

762   ophiolite peridotite slice along Indus suture zone, Ladakh Himalaya, India: Implications on melt-

763   rock interaction in a subduction-zone environment. Geochemistry, *79*(1), 78–93.

764   https://doi.org/10.1016/j.chemer.2018.10.003

765   Black, P. M., & Brothers, R. N. (1977). Blueschist ophiolites in the melange zone, northern New

766   Caledonia. Contributions to Mineralogy and Petrology, 65(1), 69–78. https://doi.org/10.1007/B

767   F00373572

768   Black, P. M. (1993). Tectonism, magmatism and sedimentary basin development, Paleozoic to

769   Paleogene, New Caledonia. Bulletin of the Geological Society of Malaysia, 33, 331–341.

770   https://doi.org/10.7186/bgsm33199323

771   Blanco-Quintero, I. F., García-Casco, A., & Gerya, T. V. (2011). Tectonic blocks in serpentinite

772   mélange (eastern Cuba) reveal large-scale convective flow of the subduction channel. Geology,

773   39(1), 79-82. https://doi.org/10.1130/G31494.1

774   Bonatti, E. (1976). Serpentinite protrusions in the oceanic crust. Earth and Planetary Science

775   Letters, *32*(2), 107-113. https://doi.org/10.1016/0012-821X(76)90048-0

776 Bostock, M. G., Hyndman, R. D., Rondenay, S., & Peacock, S. M. (2002). An inverted

777 continental Moho and serpentinization of the forearc mantle. Nature, 417(6888), 536–538.

778 https://doi.org/10.1038/417536a

779 Brothers, R. N. (1974). High-pressure schists in Northern New Caledonia. Contributions to

780 Mineralogy and Petrology, 46(2), 109–127. https://doi.org/10.1007/BF00377499

781 Brothers, R. N., & Blake, M. C. (1973). Tertiary plate tectonics and high-pressure

782 metamorphism in New Caledonia. Tectonophysics, 17(4), 337–358.

783 https://doi.org/10.1016/0040-1951(73)90046-2

784 Burkhard, D. J. M., & O'Neil, J. R. (1988). Contrasting serpentinization processes in the eastern

785 Central Alps. Contributions to Mineralogy and Petrology, 99(4), 498–506.

786 https://doi.org/10.1007/BF00371940

787 Campbell, H. J., Grant-Mackie, J. A., & Paris, J. P. (1985). Geology of the Moindou-Teremba

788 area, New Caledonia. Stratigraphy and structure of Teremba Group (Permian-Lower Triassic)

789 and Baie de St.-Vincent Group (Upper Triassic- Lower Jurassic). Géologie de la France, 1, 19-

790 36.

791 Cannaò, E., Scambelluri, M., Agostini, S., Tonarini, S., & Godard, M. (2016). Linking

792 serpentinite geochemistry with tectonic evolution at the subduction plate-interface: The Voltri

793 Massif case study (Ligurian Western Alps, Italy). Geochimica et Cosmochimica Acta, 190, 115–

794 133. https://doi.org/10.1016/j.gca.2016.06.034

795 Cannat, M. (1993). Emplacement of mantle rocks in the seafloor at mid-ocean ridges. Journal of

796 Geophysical Research: Solid Earth, 98(B3), 4163–4172. https://doi.org/10.1029/92JB02221

797 Cannat, M., Fontaine, F., & Escartín, J. (2010). Serpentinization and associated hydrogen and

798 methane fluxes at slow spreading ridges. In P. A. Rona, C. W. Devey, J. Dyment, & B. J. Murton

799 (Eds.), Diversity of Hydrothermal Systems on Slow Spreading Ocean Ridges (pp. 241–264).

800 American Geophysical Union. https://doi.org/10.1029/2008GM000760

801 Cárdenas-Párraga, J., García-Casco, A., Proenza, J. A., Harlow, G. E., Blanco-Quintero, I. F.,
802 Lázaro, C., et al. (2017). Trace-element geochemistry of transform-fault serpentinite in high-
803 pressure subduction mélanges (eastern Cuba): implications for subduction initation. International
804 Geology Review, 59(16), 2041-2064. https://doi.org/10.1080/00206814.2017.1308843

805 Clarke, G. L., Aitchison, J. C., & Cluzel, D. (1997). Eclogites and Blueschists of the Pam
806 Peninsula, NE New Caledonia: a Reappraisal. Journal of Petrology, 38(7), 843–876.
807 https://doi.org/10.1093/petroj/38.7.843

808 Cloos, M., & Shreve, R. L. (1988). Subduction-channel model of prism accretion, melange
809 formation, sediment subduction, and subduction erosion at convergent plate margins: 1.
810 Background and description. Pure and Applied Geophysics PAGEOPH, 128(3–4), 455–500.
811 https://doi.org/10.1007/BF00874548

812 Cluzel, D., Adams, C. J., Meffre, S., Campbell, H., & Maurizot, P. (2010). Discovery of Early
813 Cretaceous Rocks in New Caledonia: New Geochemical and U-Pb Zircon Age Constraints on
814 the Transition from Subduction to Marginal Breakup in the Southwest Pacific. The Journal of
815 Geology, 118(4), 381–397. https://doi.org/10.1086/652779

816 Cluzel, D. (2021). Subduction erosion: contributions of footwall and hanging wall to serpentinite
817 mélange; field, geochemical and radiochronological evidence from the Eocene HP-LT belt of
818 New Caledonia. Australian Journal of Earth Sciences, 68(1), 99-119.
819 https://doi.org/10.1080/08120099.2020.1761876

820 Cluzel, D., Aitchison, J. C., Clarke, G. L., Meffre, S., & Picard, C. (1994). Point de vue sur
821 l'évolution tectonique et géodynamique de la Nouvelle-Calédonie (Pacifique, France). Comptes
822 Rendus De L'Académie Des Sciences. Série II. Fascicule A, Sciences De La Terre Et Des
823 Planètes, 319(6), 683-690.

824 Cluzel, D., Aitchison, J. C., Clarke, G. L., Meffre, S., & Picard, C. (1995). Dénudation
825 tectonique du complexe à noyau métamorphique de haute pression d'âge tertiaire (Nord de la
826 Nouvelle-Calédonie, Pacifique, France). Comptes Rendus De L'Académie Des Sciences. Série II.
827 Fascicule A, Sciences De La Terre Et Des Planètes, 321(1), 57-64.

828   Cluzel, D., Boulvais, P., Iseppi, M., Lahondère, D., Lesimple, S., Maurizot, P., et al. (2020).
829   Slab-derived origin of tremolite–antigorite veins in a supra-subduction ophiolite: the Peridotite
830   Nappe (New Caledonia) as a case study. International Journal of Earth Sciences, 109(1), 171-
831   196. https://doi.org/10.1007/s00531-019-01796-6

832   Cluzel, D., Jourdan, F., Meffre, S., Maurizot, P., & Lesimple, S. (2012). The metamorphic sole
833   of New Caledonia ophiolite: 40 Ar/ 39 Ar, U-Pb, and geochemical evidence for subduction
834   inception at a spreading ridge. Tectonics, 31(3). https://doi.org/10.1029/2011TC003085

835   Cluzel, D., & Meffre, S. (2002). L'unité de la Boghen (Nouvelle-Calédonie, Pacifique sud-
836   ouest): Un complexe d'accrétion jurassique. Données radiochronologiques préliminaires U-Pb
837   sur les zircons détritiques. Comptes Rendus - Geoscience, 334(11), 867–874.
838   https://doi.org/10.1016/S1631-0713(02)01823-0

839   Cluzel, D., Meffre, S., Maurizot, P., & Crawford, A. J. (2006). Earliest Eocene (53 Ma)
840   convergence in the Southwest Pacific: evidence from pre-obduction dikes in the ophiolite of
841   New Caledonia. Terra Nova, 18(6), 395-402. https://doi.org/10.1111/j.1365-3121.2006.00704.x

842   Cluzel, D., Ulrich, M., Jourdan, F., Meffre, S., Paquette, J.-L., Audet, M.-A., et al. (2016). Early
843   Eocene clinoenstatite boninite and boninite-series dikes of the ophiolite of New Caledonia; a
844   witness of slab-derived enrichment of the mantle wedge in a nascent volcanic arc. Lithos, 260,
845   429-442. https://doi.org/10.1016/j.lithos.2016.04.031

846   Codillo, E. A., Le Roux, V., & Marschall, H. R. (2018). Arc-like magmas generated by mélange-
847   peridotite interaction in the mantle wedge. Nature Communications, 9(1), 2864.
848   https://doi.org/10.1038/s41467-018-05313-2

849   Coleman, R. G. (1967). Glaucophane schists from California and New Caledonia.
850   Tectonophysics, 4(4–6), 479–498. https://doi.org/10.1016/0040-1951(67)90012-1

851   Collins, N. C., Bebout, G. E., Angiboust, S., Agard, P., Scambelluri, M., Crispini, L., & John, T.
852   (2015). Subduction zone metamorphic pathway for deep carbon cycling: II. Evidence from

853   HP/UHP metabasaltic rocks and ophicarbonates. Chemical Geology, 412, 132-150.

854   https://doi.org/10.1016/j.chemgeo.2015.06.012

855   Collot, J. Y., Malahoff, A., Recy, J., Latham, G., & Missegue, F. (1987). Overthrust

856   emplacement of New Caledonia Ophiolite: Geophysical evidence. Tectonics, 6(3), 215–232.

857   https://doi.org/10.1029/TC006i003p00215

858   Cooperdock, E. H. G., Raia, N. H., Barnes, J. D., Stockli, D. F., & Schwarzenbach, E. M. (2018).

859   Tectonic origin of serpentinites on Syros, Greece: Geochemical signatures of abyssal origin

860   preserved in a HP/LT subduction complex. Lithos, 296–299(January), 352–364.

861   https://doi.org/10.1016/j.lithos.2017.10.020

862   Cooperdock, E. H. G., & Stockli, D. F. (2016). Unraveling alteration histories in serpentinites

863   and associated ultramafic rocks with magnetite (U-Th)/He geochronology. Geology, 44(11),

864   967–970. https://doi.org/10.1130/G38587.1

865   Crawford, A. J., Meffre, S., & Symonds, P. A. (2003). 120 to 0 Ma tectonic evolution of the

866   southwest Pacific and analogous geological evolution of the 600 to 220 Ma Tasman Fold Belt

867   System. In Evolution and Dynamics of the Australian Plate, 22, 383-403.

868   https://doi.org/10.1130/0-8137-2372-8.383

869   Davies, H. L., & Smith, I. E. (1971). Geology of eastern Papua. Geological Society of America

870   Bulletin, 82(12), 3299-3312.

871   Debret, B., Albers, E., Walter, B., Price, R., Barnes, J. D., Beunon, H., et al. (2019). Shallow

872   forearc mantle dynamics and geochemistry: New insights from IODP Expedition 366. Lithos,

873   326-327, 230-245. https://doi.org/10.1016/j.lithos.2018.10.038

874   Debret, B., Reekie, C. D. J., Mattielli, N., Beunon, H., Ménez, B., Savov, I., & Williams, H. M.

875   (2020). Redox transfer at subduction zones: insights from Fe isotopes in the Mariana forearc.

876   Geochemical Perspective Letters, 46-51. https://doi.org/10.7185/geochemlet.2003

877   Debret, B., & Sverjensky, D. A. (2017). Highly oxidising fluids generated during serpentinite

878   breakdown in subduction zones. Scientific Reports, 7(1), 10351. https://doi.org/10.1038/s41598-

879   017-09626-y

880   Deschamps, F., Godard, M., Guillot, S., & Hattori, K. (2013). Geochemistry of subduction zone

881   serpentinites: A review. Lithos, 178, 96–127. https://doi.org/10.1016/j.lithos.2013.05.019

882   Deschamps, F., Guillot, S., Godard, M., Andreani, M., & Hattori, K. (2011). Serpentinites act as

883   sponges for fluid-mobile elements in abyssal and subduction zone environments. Terra Nova,

884   23(3), 171-178. https://doi.org/10.1111/j.1365-3121.2011.00995.x

885   Deschamps, F., Guillot, S., Godard, M., Chauvel, C., Andreani, M., & Hattori, K. (2010). In situ

886   characterization of serpentinites from forearc mantle wedges: Timing of serpentinization and

887   behavior of fluid-mobile elements in subduction zones. Chemical Geology, 269(3-4), 262-277.

888   https://doi.org/10.1016/j.chemgeo.2009.10.002

889   Dubois, J., Launay, J., & Recy, J. (1974). Uplift movements in New Caledonia-Loyalty Islands

890   area and their plate tectonics interpretation. Tectonophysics, 24(1–2), 133–150.

891   https://doi.org/10.1016/0040-1951(74)90134-6

892   Dupuy, C., Dostal, J., & Leblanc, M. (1981). Geochemistry of an ophiolitic complex from New

893   Caledonia. Contributions to Mineralogy and Petrology, 76(1), 77–83.

894   https://doi.org/10.1007/BF00373686

895   Eiler, J. M. (2001). 5. Oxygen Isotope Variations of Basaltic Lavas and Upper Mantle Rocks. In

896   Stable Isotope Geochemistry (Vol. 43, pp. 319–364). De Gruyter.

897   https://doi.org/10.1515/9781501508745-008

898   Evans, K. A. (2012). The redox budget of subduction zones. Earth-Science Reviews, 113(1–2),

899   11–32. https://doi.org/10.1016/j.earscirev.2012.03.003

900   Evans, K. A., Reddy, S. M., Tomkins, A. G., Crossley, R. J., & Frost, B. R. (2017). Effects of

901   geodynamic setting on the redox state of fluids released by subducted mantle lithosphere. Lithos,

902   278-281, 26-42. https://doi.org/10.1016/j.lithos.2016.12.023

903   Evans, K. A., & Frost, B. R. (2021). Deserpentinization in Subduction Zones as a Source of

904   Oxidation in Arcs: a Reality Check. Journal of Petrology, 62(3), 1–32.

905   https://doi.org/10.1093/petrology/egab016

906   Ferrand, T. P. (2019). Seismicity and mineral destabilizations in the subducting mantle up to 6

907   GPa, 200 km depth. Lithos, 334–335, 205–230. https://doi.org/10.1016/j.lithos.2019.03.014

908   Fitzherbert, J. A., Clarke, G. L., & Powell, R. (2003). Lawsonite-Omphacite-Bearing

909   Metabasites of the Pam Peninsula, NE New Caledonia: Evidence for Disrupted Blueschist- to

910   Eclogite-Facies Conditions. Journal of Petrology, 44(10), 1805-1831.

911   https://doi.org/10.1093/petrology/egg060

912   Fitzherbert, J. A., Clarke, G. L., Marmo, B., & Powell, R. (2004). The origin and P-T evolution

913   of peridotites and serpentinites of NE New Caledonia: prograde interaction between continental

914   margin and the mantle wedge. Journal of Metamorphic Geology, 22(4), 327-344.

915   https://doi.org/10.1111/j.1525-1314.2004.00517.x

916   Früh-Green, G. L., Scambelluri, M., & Vallis, F. (2001). O–H isotope ratios of high pressure

917   ultramafic rocks: implications for fluid sources and mobility in the subducted hydrous mantle.

918   Contributions to Mineralogy and Petrology, 141(2), 145-159.

919   https://doi.org/10.1007/s004100000228

920   Früh-Green, G. L., Weissert, H., & Bernoulli, D. (1990). A multiple fluid history recorded in

921   Alpine ophiolites. Journal of the Geological Society, 147(6), 959–970.

922   https://doi.org/10.1144/gsjgs.147.6.0959

923   Fyfe, W., & McBirney, A. (1975). Subduction and the structure of andesitic volcanic belts.

924   American Journal of Science, Vol. 275-A, pp. 285–297.

925   Gautier, P., Quesnel, B., Boulvais, P., & Cathelineau, M. (2016). The emplacement of the

926   Peridotite Nappe of New Caledonia and its bearing on the tectonics of obduction. Tectonics,

927   35(12), 3070–3094. https://doi.org/10.1002/2016TC004318

928   Gerya, T. V., Stöckhert, B., & Perchuk, A. L. (2002). Exhumation of high-pressure metamorphic

929   rocks in a subduction channel: A numerical simulation. Tectonics, 21(6), 6-1-6–19.

930   https://doi.org/10.1029/2002TC001406

931   Gerya, T., & Stöckhert, B. (2006). Two-dimensional numerical modeling of tectonic and

932   metamorphic histories at active continental margins. International Journal of Earth Sciences,

933   95(2), 250–274. https://doi.org/10.1007/s00531-005-0035-9

934   Groppo, C., Rinaudo, C., Cairo, S., Gastaldi, D., & Compagnoni, R. (2006). Micro-Raman

935   spectroscopy for a quick and reliable identification of serpentine minerals from ultramafics.

936   European Journal of Mineralogy, 18(3), 319-329. https://doi.org/10.1127/0935-1221/2006/0018-

937   0319

938   Guérangé, B., Lozes, J., Autran, A., 1977. Mesozoic metamorphism in the New Caledonia

939   central chain and its geodynamic implications in relation to the evolution of the Cretaceous

940   Rangitata Orogeny. In: International Symposium on Geodynamics in South-West Pacific (pp.

941   265–278). Paris: Technip.

942   Guice, G. L., Ackerson, M. R., Holder, R. M., George, F. R., Browning-Hanson, J. F., Burgess,

943   J. L., et al. (2021). Suprasubduction zone ophiolite fragments in the central Appalachian orogen:

944   Evidence for mantle and Moho in the Baltimore Mafic Complex (Maryland, USA). Geosphere,

945   17(2), 561-581. https://doi.org/10.1130/GES02289.1

946   Guillot, S., Hattori, K. H., & de Sigoyer, J. (2000). Mantle wedge serpentinization and

947   exhumation of eclogites: Insights from eastern Ladakh, northwest Himalaya. Geology, 28(3),

948   199–202. https://doi.org/10.1130/0091-7613(2000)028<0199:MWSAEO>2.3.CO;2

949   Guillot, S., Hattori, K., Agard, P., Schwartz, S., & Vidal, O. (2009). Exhumation Processes in

950   Oceanic and Continental Subduction Contexts: A Review. In: Lallemand S., Funiciello F. (eds)

951   Subduction Zone Geodynamics. Frontiers in Earth Sciences. Springer, Berlin, Heidelberg.

952   https://doi.org/10.1007/978-3-540-87974-9_10

953   Guillot, S., Schwartz, S., Reynard, B., Agard, P., & Prigent, C. (2015). Tectonic significance of

954   serpentinites. Tectonophysics, 646(February), 1–19. https://doi.org/10.1016/j.tecto.2015.01.020

955   Hacker, B. R., Peacock, S. M., Abers, G. A., & Holloway, S. D. (2003). Subduction factory 2.

956   Are intermediate-depth earthquakes in subducting slabs linked to metamorphic dehydration

957   reactions? Journal of Geophysical Research: Solid Earth, 108(B1).

958   https://doi.org/10.1029/2001JB001129

959   Hacker, B. R. (2008). H2O subduction beyond arcs. Geochemistry, Geophysics, Geosystems,

960   9(3), Q03001. https://doi.org/10.1029/2007GC001707

961   Halama, R., Bebout, G. E., John, T., & Scambelluri, M. (2014). Nitrogen recycling in subducted

962   mantle rocks and implications for the global nitrogen cycle. International Journal of Earth

963   Sciences, 103(7), 2081–2099. https://doi.org/10.1007/s00531-012-0782-3

964   Hart, S. R., & Zindler, A. (1986). In search of a bulk-Earth composition. Chemical Geology,

965   57(3–4), 247–267. https://doi.org/10.1016/0009-2541(86)90053-7

966   Hattori, K. H., & Guillot, S. (2003). Volcanic fronts form as a consequence of serpentinite

967   dehydration in the forearc mantle wedge. Geology, 31(6), 525. https://doi.org/10.1130/0091-

968   7613(2003)031<0525:VFFAAC>2.0.CO;2

969   Hattori, K., Wallis, S., Enami, M., & Mizukami, T. (2010). Subduction of mantle wedge

970   peridotites: Evidence from the Higashi-akaishi ultramafic body in the Sanbagawa metamorphic

971   belt. Island Arc, 19(1), 192-207. https://doi.org/10.1111/j.1440-1738.2009.00696.x

972   Hermann, J., Müntener, O., & Scambelluri, M. (2000). The importance of serpentinite mylonites

973   for subduction and exhumation of oceanic crust. Tectonophysics, 327(3–4), 225–238.

974   https://doi.org/10.1016/S0040-1951(00)00171-2

975   Hilairet, N., Daniel, I., & Reynard, B. (2006). Equation of state of antigorite, stability field of

976   serpentines, and seismicity in subduction zones. Geophysical Research Letters, 33(2), L02302.

977   https://doi.org/10.1029/2005GL024728

978   Hilairet, N., & Reynard, B. (2009). Stability and dynamics of serpentinite layer in subduction

979   zone. Tectonophysics, 465(1–4), 24–29. https://doi.org/10.1016/j.tecto.2008.10.005

980   Hilairet, N., Reynard, B., Wang, Y., Daniel, I., Merkel, S., Nishiyama, N., & Petitgirard, S.

981   (2007). High-Pressure Creep of Serpentine, Interseismic Deformation, and Initiation of

982   Subduction. Science, 318(5858), 1910–1913. https://doi.org/10.1126/science.1148494

983   Hirth, G., & Guillot, S. (2013). Rheology and Tectonic Significance of Serpentinite. Elements,

984   9(2), 107–113. https://doi.org/10.2113/gselements.9.2.107

985   Hyndman, R. D., & Peacock, S. M. (2003). Serpentinization of the forearc mantle. Earth and

986   Planetary Science Letters, 212(3–4), 417–432. https://doi.org/10.1016/S0012-821X(03)00263-2

987   Jagoutz, E., Palme, H., Baddenhausen, H., Blum, K., Cendales, M., Dreibus, G., et al. (1979).

988   The abundance of major, minor and trace elements in the earth's mantle as derived from

989   primitive ultramafic nodules. Lunar and Planetary Science Conference Proceedings, 10, 2031–

990   2050.

991   John, T., Scambelluri, M., Frische, M., Barnes, J. D., & Bach, W. (2011). Dehydration of

992   subducting serpentinite: Implications for halogen mobility in subduction zones and the deep

993   halogen cycle. Earth and Planetary Science Letters, 308(1–2), 65–76.

994   https://doi.org/10.1016/j.epsl.2011.05.038

995   Johnson, D., Hooper, P., & Conrey, R. (1999). XRF method XRF analysis of rocks and minerals

996   for major and trace elements on a single low dilution Li-tetraborate fused bead. Advances in X-

997   ray Analysis, 41, 843-867.

998   Jung, H., & Green, H. W. (2004). Experimental Faulting of Serpentinite during Dehydration:

999   Implications for Earthquakes, Seismic Low-Velocity Zones, and Anomalous Hypocenter

1000   Distributions in Subduction Zones. International Geology Review, 46(12), 1089-1102.

1001   https://doi.org/10.2747/0020-6814.46.12.1089

1002   Katzir, Y., Avigad, D., Matthews, A., Garfunkel, Z., & Evans, B. W. (2000). Origin, HP/LT

1003   metamorphism and cooling of ophiolitic mélanges in southern Evia (NW Cyclades), Greece.

1004   Journal of Metamorphic Geology, 18(6), 699-718. https://doi.org/10.1046/j.1525-
1005   1314.2000.00281.x

1006   Kelly, D. S. (2018). Analysis of Geological Materials by Low Dilution Fusion at the Peter
1007   Hooper GeoAnalytical Lab (Washington State University). Washington State University,
1008   Pullman, WA, USA.

1009   Kendrick, M. A., Hémond, C., Kamenetsky, V. S., Danyushevsky, L., Devey, C. W., Rodemann,
1010   T., et al. (2017). Seawater cycled throughout Earth's mantle in partially serpentinized
1011   lithosphere. Nature Geoscience, 10(3), 222-228. https://doi.org/10.1038/ngeo2902

1012   Kendrick, M. A., Scambelluri, M., Honda, M., & Phillips, D. (2011). High abundances of noble
1013   gas and chlorine delivered to the mantle by serpentinite subduction. Nature Geoscience, 4(11),
1014   807-812. https://doi.org/10.1038/ngeo1270

1015   Kerrick, D. M., & Connolly, J. A. D. (1998). Subduction of ophicarbonates and recycling of CO_2
1016   and H_2O. Geology, *26*(4), 375-378. https://doi.org/10.1130/0091-
1017   7613(1998)026<0375:SOOARO>2.3.CO;2

1018   Kerswell, B. C., Kohn, M. J., & Gerya, T. V. (2021). Backarc Lithospheric Thickness and
1019   Serpentine Stability Control Slab-Mantle Coupling Depths in Subduction Zones. Geochemistry,
1020   Geophysics, Geosystems, 22(6). https://doi.org/10.1029/2020GC009304

1021   Klingelhoefer, F., Lafoy, Y., Collot, J., Cosquer, E., Géli, L., Nouzé, H., & Vially, R. (2007).
1022   Crustal structure of the basin and ridge system west of New Caledonia (southwest Pacific) from
1023   wide-angle and reflection seismic data. Journal of Geophysical Research, 112(B11), B11102.
1024   https://doi.org/10.1029/2007JB005093

1025   Knaack, C. M., Cornelius, S., & Hooper, P. R. (1994). Trace element analysis of rocks and
1026   minerals by ICP–MS: Washington State University. Geology Department, Open File Report.

1027   Kodolányi, J., Pettke, T., Spandler, C., Kamber, B. S., & Gméling, K. (2012). Geochemistry of
1028   Ocean Floor and Fore-arc Serpentinites: Constraints on the Ultramafic Input to Subduction
1029   Zones. Journal of Petrology, 53(2), 235-270. https://doi.org/10.1093/petrology/egr058

1030   Kyser, T. K., O'Hanley, D. S., & Wicks, F. J. (1999). The origin of fluids associated with

1031   serpentinization processes: Evidence from stable-isotope compositions. Canadian Mineralogist,

1032   37(1), 223–237.

1033   Kyser, T. K., & O'Neil, J. R. (1984). Hydrogen isotope systematics of submarine basalts.

1034   Geochimica et Cosmochimica Acta, 48(10), 2123–2133. https://doi.org/10.1016/0016-

1035   7037(84)90392-2

1036   Lafoy, Y., Brodien, I., Vially, R., & Exon, N. F. (2005). Structure of the Basin and Ridge System

1037   West of New Caledonia (Southwest Pacific): A Synthesis. Marine Geophysical Researches,

1038   26(1), 37-50. https://doi.org/10.1007/s11001-005-5184-5

1039   Lazar, C., Cooperdock, E. H. G., & Seymour, B. H. T. (2021). A continental forearc serpentinite

1040   diapir with deep origins: Elemental signatures of a mantle wedge protolith and slab-derived

1041   fluids at New Idria, California. Lithos, 398-399, 106252.

1042   https://doi.org/10.1016/j.lithos.2021.106252

1043   Li, H.-Y., Chen, R.-X., Zheng, Y.-F., Hu, Z., & Xu, L. (2018). Crustal Metasomatism at the

1044   Slab-Mantle Interface in a Continental Subduction Channel: Geochemical Evidence From

1045   Orogenic Peridotite in the Sulu Orogen. Journal of Geophysical Research: Solid Earth, 123(3),

1046   2174–2198. https://doi.org/10.1002/2017JB014015

1047   Li, X.-P., Rahn, M., & Bucher, K. (2004). Serpentinites of the Zermatt-Saas ophiolite complex

1048   and their texture evolution. Journal of Metamorphic Geology, 22(3), 159–177.

1049   https://doi.org/10.1111/j.1525-1314.2004.00503.x

1050   Lillie, A. (1975). Structures in the lawsonite—glaucophane schists of New

1051   Caledonia. Geological Magazine, *112*(3), 225-240. https://doi.org/10.1017/S0016756800046999

1052   Liu, C.-Z., Xu, Y., & Wu, F.-Y. (2018). Limited recycling of crustal osmium in forearc mantle

1053   during slab dehydration. Geology, 46(3), 239–242. https://doi.org/10.1130/G39869.1

1054   Magott, R., Fabbri, O., & Fournier, M. (2020). Seismically-induced serpentine dehydration as a

1055   possible mechanism of water release in subduction zones. Insights from the Alpine Corsica

1056   pseudotachylyte-bearing Monte Maggiore ophiolitic unit. Lithos, 362-363, 105474.

1057   https://doi.org/10.1016/j.lithos.2020.105474

1058   Malatesta, C., Gerya, T., Scambelluri, M., Federico, L., Crispini, L., & Capponi, G. (2012).

1059   Intraoceanic subduction of "heterogeneous" oceanic lithosphere in narrow basins: 2D numerical

1060   modeling. Lithos, 140–141, 234–251. https://doi.org/10.1016/j.lithos.2012.01.003

1061   Malvoisin, B. (2015). Mass transfer in the oceanic lithosphere: Serpentinization is not

1062   isochemical. Earth and Planetary Science Letters, 430, 75–85.

1063   https://doi.org/10.1016/j.epsl.2015.07.043

1064   Marchesi, C., Garrido, C. J., Godard, M., Belley, F., & Ferré, E. (2009). Migration and

1065   accumulation of ultra-depleted subduction-related melts in the Massif du Sud ophiolite (New

1066   Caledonia). Chemical Geology, 266(3-4), 171-186.

1067   https://doi.org/10.1016/j.chemgeo.2009.06.004

1068   Marschall, H. R., & Schumacher, J. C. (2012). Arc magmas sourced from mélange diapirs in

1069   subduction zones. Nature Geoscience, 5(12), 862–867. https://doi.org/10.1038/ngeo1634

1070   Mattey, D., Lowry, D., & Macpherson, C. (1994). Oxygen isotope composition of mantle

1071   peridotite. Earth and Planetary Science Letters, 128(3–4), 231–241.

1072   https://doi.org/10.1016/0012-821X(94)90147-3

1073   Maurizot, P. (2011). First sedimentary record of the pre-obduction convergence in New

1074   Caledonia: formation of an Early Eocene accretionary complex in the north of Grande Terre and

1075   emplacement of the 'Montagnes Blanches' nappe. Bulletin de la Société Géologique de France,

1076   182(6), 479-491. https://doi.org/10.2113/gssgfbull.182.6.479

1077   Maurizot, P. & Collot, J. (2009). Explanatory note of the geological map of New Caledonia,

1078   scale 1/500,000. Direction de l'Industrie, des Mines et de l'Energie – Service géologique de

1079   Nouvelle Calédonie, Bureau de Recherches Géologiques et Minières.

1080 Maurizot, P., Eberle, J-M, Habault, C. & Tessarollo, C. (1989). Carte géologique á l'échelle du
1081 l/50,000 et notice explicative sur la feuille Pam-Ouégoa, 2nd edn. Nouméa, New Caledonia:
1082 Bureau de Recherches Geologique et Mineres.

1083 Maurizot, P., Robineau, B., Vendé-Leclerc, M., & Cluzel, D. (2020a). Introduction to New
1084 Caledonia: geology, geodynamic evolution and mineral resources. Geological Society, London,
1085 Memoirs, 51(1), 1-12. https://doi.org/10.1144/M51

1086 Maurizot, P., Cluzel, D., Meffre, S., Campbell, H. J., Collot, J., & Sevin, B. (2020b). Chapter 3
1087 Pre-Late Cretaceous basement terranes of the Gondwana active margin of New Caledonia.
1088 Geological Society,London,Memoirs, 51(1), 27-52. https://doi.org/10.1144/M51-2016-11

1089 Maurizot, P., Cluzel, D., Patriat, M., Collot, J., Iseppi, M., Lesimple, S., et al. (2020c). Chapter 5
1090 The Eocene Subduction–Obduction Complex of New Caledonia. Geological Society, London,
1091 Memoirs, 51(1), 93-130. https://doi.org/10.1144/M51-2018-70

1092 McDonough, W. F., & Sun, S. -s. (1995). The composition of the Earth. Chemical Geology,
1093 120(3–4), 223–253. https://doi.org/10.1016/0009-2541(94)00140-4

1094 Meffre, S., Aitchison, J. C., & Crawford, A. J. (1996). Geochemical stratigraphy of boninites and
1095 tholeiites from the Permo-Triassic Koh Ophiolite, New Caledonia. Tectonics, 15(1), 67–83.

1096 Mével, C. (2003). Serpentinization of abyssal peridotites at mid-ocean ridges. Comptes Rendus
1097 Geoscience, 335(10–11), 825–852. https://doi.org/10.1016/j.crte.2003.08.006

1098 Mortimer, N., Gans, P. B., Meffre, S., Martin, C. E., Seton, M., Williams, S., … Rollet, N.
1099 (2018). Regional volcanism of northern Zealandia: post-Gondwana break-up magmatism on an
1100 extended, submerged continent. Geological Society, London, Special Publications, 463(1), 199-
1101 226. https://doi.org/10.1144/SP463.9

1102 Mortimer, N., Campbell, H. J., Tulloch, A. J., King, P. R., Stagpoole, V. M., Wood, R. A., …
1103 Seton, M. (2017). Zealandia: Earth's Hidden Continent. GSA Today, 27(3–4), 27–35.
1104 https://doi.org/10.1130/GSATG321A.1

1105   Mothersole, F. E. (2014). A Comparative Study of Progressive Serpentinisation. Curtin
1106   University.

1107   Mothersole, F. E., Evans, K., & Frost, B. R. (2017). Abyssal and hydrated mantle wedge
1108   serpentinised peridotites: A comparison of the 15∘20′N fracture zone and New Caledonia
1109   serpentinites. Contributions to Mineralogy and Petrology, 172(8), 69.
1110   https://doi.org/10.1007/s00410-017-1381-x

1111   Moutte, J. (1982). Chromite deposits of the Tiebaghi ultramafic massif, New Caledonia.
1112   Economic Geology, 77(3), 576–591. https://doi.org/10.2113/gsecongeo.77.3.576

1113   Nielsen, S. G., & Marschall, H. R. (2017). Geochemical evidence for mélange melting in global
1114   arcs. Science Advances, 3(4), 1–7. https://doi.org/10.1126/sciadv.1602402

1115   Niu, Y. (2004). Bulk-rock Major and Trace Element Compositions of Abyssal Peridotites:
1116   Implications for Mantle Melting, Melt Extraction and Post-melting Processes Beneath Mid-
1117   Ocean Ridges. Journal of Petrology, 45(12), 2423-2458.
1118   https://doi.org/10.1093/petrology/egh068

1119   Paris, J. P. (1981). Géologie de la Nouvelle-Calédonie, un essai de synthèse. Mémoires du
1120   Bureau de Recherches Géologiques et Minières, 113.

1121   Parkinson, I. J., & Pearce, J. A. (1998). Peridotites from the Izu-Bonin-Mariana Forearc (ODP
1122   Leg 125): Evidence for Mantle Melting and Melt-Mantle Interaction in a Supra-Subduction Zone
1123   Setting. Journal of Petrology, 39(9), 1577–1618. https://doi.org/10.1093/petroj/39.9.1577

1124   Patriat, M., Collot, J., Etienne, S., Poli, S., Clerc, C., Mortimer, N., … Roest, W. R. (2018). New
1125   Caledonia Obducted Peridotite Nappe: Offshore Extent and Implications for Obduction and
1126   Postobduction Processes. Tectonics, 37(4), 1077-1096. https://doi.org/10.1002/2017TC004722

1127   Peacock, S. M., & Hyndman, R. D. (1999). Hydrous minerals in the mantle wedge and the
1128   maximum depth of subduction thrust earthquakes. Geophysical Research Letters, 26(16), 2517–
1129   2520. https://doi.org/10.1029/1999GL900558

1130   Peacock, S. M. (2001). Are the lower planes of double seismic zones caused by serpentine

1131   dehydration in subducting oceanic mantle? Geology, 29(4), 299. https://doi.org/10.1130/0091-

1132   7613(2001)029<0299:ATLPOD>2.0.CO;2

1133   Peacock, S. M., & Wang, K. (2021). On the Stability of Talc in Subduction Zones: A Possible

1134   Control on the Maximum Depth of Decoupling Between the Subducting Plate and Mantle

1135   Wedge. Geophysical Research Letters, 48(17), 1–8. https://doi.org/10.1029/2021GL094889

1136   Peters, D., Bretscher, A., John, T., Scambelluri, M., & Pettke, T. (2017). Fluid-mobile elements

1137   in serpentinites: Constraints on serpentinisation environments and element cycling in subduction

1138   zones. Chemical Geology, 466(March), 654–666. https://doi.org/10.1016/j.chemgeo.2017.07.017

1139   Petriglieri, J. R., Salvioli-Mariani, E., Mantovani, L., Tribaudino, M., Lottici, P. P., Laporte-

1140   Magoni, C., & Bersani, D. (2015). Micro-Raman mapping of the polymorphs of serpentine.

1141   Journal of Raman Spectroscopy, 46(10), 953-958. https://doi.org/10.1002/jrs.4695

1142   Pirard, C., & Hermann, J. (2015). Focused fluid transfer through the mantle above subduction

1143   zones. Geology, 43(10), 915–918. https://doi.org/10.1130/G37026.1

1144   Pirard, C., Hermann, J., & O'Neill, H. S. C. (2013). Petrology and Geochemistry of the Crust–

1145   Mantle Boundary in a Nascent Arc, Massif du Sud Ophiolite, New Caledonia, SW Pacific.

1146   Journal of Petrology, 54(9), 1759-1792. https://doi.org/10.1093/petrology/egt030

1147   Pirard, C., & Spandler, C. (2017). The zircon record of high-pressure metasedimentary rocks of

1148   New Caledonia: Implications for regional tectonics of the south-west Pacific. Gondwana

1149   Research, 46, 79-94. https://doi.org/10.1016/j.gr.2017.03.001

1150   Potel, S., Mählmann, R. F., Stern, W. B., Mullis, J., & Frey, M. (2006). Very Low-grade

1151   Metamorphic Evolution of Pelitic Rocks under High-pressure/Low-temperature Conditions, NW

1152   New Caledonia (SW Pacific). Journal of Petrology, 47(5), 991–1015.

1153   https://doi.org/10.1093/petrology/egl001

1154   Prinzhofer, A., 1981. Structure et petrologie d'un cortège ophiolitique: le massif du Sud

1155   (Nouvelle Calédonie). Ph.D. thesis. Ecole Nationale Supérieure des Mines de Paris.

1156   Prinzhofer, A., & Nicolas, A. (1980). The Bogota Peninsula, New Caledonia: A Possible

1157   Oceanic Transform Fault. The Journal of Geology, 88(4), 387–398.

1158   https://doi.org/10.1086/628523

1159   Prinzhofer, A., & Allègre, C. J. (1985). Residual peridotites and the mechanisms of partial

1160   melting. Earth and Planetary Science Letters, 74(2–3), 251–265. https://doi.org/10.1016/0012-

1161   821X(85)90025-1

1162   Proctor, B., & Hirth, G. (2015). Role of pore fluid pressure on transient strength changes and

1163   fabric development during serpentine dehydration at mantle conditions: Implications for

1164   subduction-zone seismicity. Earth and Planetary Science Letters, 421, 1-12.

1165   https://doi.org/10.1016/j.epsl.2015.03.040

1166   Ranero, C. R., Phipps Morgan, J., McIntosh, K., & Reichert, C. (2003). Bending-related faulting

1167   and mantle serpentinization at the Middle America trench. Nature, 425(6956), 367–373.

1168   https://doi.org/10.1038/nature01961

1169   Rawling, T. J., & Lister, G. S. (2002). Large-scale structure of the eclogite–blueschist belt of

1170   New Caledonia. Journal of Structural Geology, 24(8), 1239–1258.

1171   https://doi.org/10.1016/S0191-8141(01)00128-6

1172   Reynard, B. (2013). Serpentine in active subduction zones. Lithos, 178, 171–185.

1173   https://doi.org/10.1016/j.lithos.2012.10.012

1174   Ribeiro, J. M., & Lee, C.-T. A. (2017). An imbalance in the deep water cycle at subduction

1175   zones: The potential importance of the fore-arc mantle. Earth and Planetary Science Letters, 479,

1176   298–309. https://doi.org/10.1016/j.epsl.2017.09.018

1177   Rouméjon, S., Cannat, M., Agrinier, P., Godard, M., & Andreani, M. (2015). Serpentinization

1178   and Fluid Pathways in Tectonically Exhumed Peridotites from the Southwest Indian Ridge (62-

1179   65 E). Journal of Petrology, 56(4), 703-734. https://doi.org/10.1093/petrology/egv014

1180   Rüpke, L. H., Morgan, J. P., Hort, M., & Connolly, J. A. (2004). Serpentine and the subduction

1181   zone water cycle. Earth and Planetary Science Letters, 223(1–2), 17–34.

1182   https://doi.org/10.1016/j.epsl.2004.04.018

1183   Saccocia, P. J., Seewald, J. S., & Shanks, W. C. (2009). Oxygen and hydrogen isotope

1184   fractionation in serpentine–water and talc–water systems from 250 to 450°C, 50MPa.

1185   Geochimica et Cosmochimica Acta, 73(22), 6789-6804.

1186   https://doi.org/10.1016/j.gca.2009.07.036

1187   Salters, V. J. M., & Stracke, A. (2004). Composition of the depleted mantle. Geochemistry,

1188   Geophysics, Geosystems, 5(5). https://doi.org/10.1029/2003GC000597

1189   Savov, I. P., Ryan, J. G., D'Antonio, M., & Fryer, P. (2007). Shallow slab fluid release across

1190   and along the Mariana arc-basin system: Insights from geochemistry of serpentinized peridotites

1191   from the Mariana fore arc. Journal of Geophysical Research, 112(B9), B09205.

1192   https://doi.org/10.1029/2006JB004749

1193   Savov, I. P., Ryan, J. G., D'Antonio, M., Kelley, K., & Mattie, P. (2005). Geochemistry of

1194   serpentinized peridotites from the Mariana Forearc Conical Seamount, ODP Leg 125:

1195   Implications for the elemental at subduction zones. Geochemistry, Geophysics, Geosystems,

1196   6(4). https://doi.org/10.1029/2004GC000777

1197   Scambelluri, M., Strating, E. H. H., Piccaardo, G. B., Vissers, R. L. M., & Rampone, E. (1991).

1198   Alpine olivine- and titanian clinohumite-bearing assemblages in the Erro-Tobbio peridotite

1199   (Voltri Massif, NW Italy). Journal of Metamorphic Geology, 9(1), 79-91.

1200   https://doi.org/10.1111/j.1525-1314.1991.tb00505.x

1201   Scambelluri, M., Cannaò, E., & Gilio, M. (2019). The water and fluid-mobile element cycles

1202   during serpentinite subduction. A review. European Journal of Mineralogy, 31(3), 405–428.

1203   https://doi.org/10.1127/ejm/2019/0031-2842

1204 Scambelluri, M., Fiebig, J., Malaspina, N., Müntener, O., & Pettke, T. (2004). Serpentinite
1205 Subduction: Implications for Fluid Processes and Trace-Element Recycling. International
1206 Geology Review, 46(7), 595–613. https://doi.org/10.2747/0020-6814.46.7.595

1207 Scambelluri, M., & Tonarini, S. (2012). Boron isotope evidence for shallow fluid transfer across
1208 subduction zones by serpentinized mantle. Geology, 40(10), 907–910.
1209 https://doi.org/10.1130/G33233.1

1210 Schmidt, M. W., & Poli, S. (1998). Experimentally based water budgets for dehydrating slabs
1211 and consequences for arc magma generation. Earth and Planetary Science Letters, 163(1–4),
1212 361–379. https://doi.org/10.1016/S0012-821X(98)00142-3

1213 Schwartz, S., Allemand, P., & Guillot, S. (2001). Numerical model of the effect of serpentinites
1214 on the exhumation of eclogitic rocks: insights from the Monviso ophiolitic massif (Western
1215 Alps). Tectonophysics, 342(1-2), 193-206. https://doi.org/10.1016/S0040-1951(01)00162-7

1216 Schwartz, S., Gautheron, C., Ketcham, R. A., Brunet, F., Corre, M., Agranier, A., et al. (2020).
1217 Unraveling the exhumation history of high-pressure ophiolites using magnetite (U-Th-Sm)/He
1218 thermochronometry. Earth and Planetary Science Letters, 543, 116359.
1219 https://doi.org/10.1016/j.epsl.2020.116359

1220 Secchiari, A., Gleissner, P., Li, C., Goncharov, A., Milke, R., Becker, H., et al. (2020). Highly
1221 siderophile and chalcophile element behaviour in abyssal-type and supra-subduction zone
1222 mantle: New insights from the New Caledonia ophiolite. Lithos, 354–355(December 2019),
1223 105338. https://doi.org/10.1016/j.lithos.2019.105338

1224 Secchiari, A., Montanini, A., Bosch, D., Macera, P., & Cluzel, D. (2016). Melt extraction and
1225 enrichment processes in the New Caledonia lherzolites: Evidence from geochemical and Sr–Nd
1226 isotope data. Lithos, 260(May), 28–43. https://doi.org/10.1016/j.lithos.2016.04.030

1227 Secchiari, A., Montanini, A., Bosch, D., Macera, P., & Cluzel, D. (2018). The contrasting
1228 geochemical message from the New Caledonia gabbronorites: insights on depletion and

1229 contamination processes of the sub-arc mantle in a nascent arc setting. Contributions to

1230 Mineralogy and Petrology, 173(8), 66. https://doi.org/10.1007/s00410-018-1496-8

1231 Sécher, D. (1981). Les lherzolites ophiolitiques de Nouvelle-Calédonie et leurs gisements de

1232 chromite. L'Institut des Sciences de la Nature de L'Université de Nantes.

1233 Sharp, Z. D., Atudorei, V., & Durakiewicz, T. (2001). A rapid method for determination of

1234 hydrogen and oxygen isotope ratios from water and hydrous minerals. Chemical Geology,

1235 178(1–4), 197–210. https://doi.org/10.1016/S0009-2541(01)00262-5

1236 Sharp, Z. D. (1990). A laser-based microanalytical method for the in situ determination of

1237 oxygen isotope ratios of silicates and oxides. Geochimica et Cosmochimica Acta, 54(5), 1353–

1238 1357. https://doi.org/10.1016/0016-7037(90)90160-M

1239 Shen, T., Hermann, J., Zhang, L., Lü, Z., Padrón-Navarta, J. A., Xia, B., & Bader, T. (2015).

1240 UHP Metamorphism Documented in Ti-chondrodite- and Ti-clinohumite-bearing Serpentinized

1241 Ultramafic Rocks from Chinese Southwestern Tianshan. Journal of Petrology, 56(7), 1425-1458.

1242 https://doi.org/10.1093/petrology/egv042

1243 Shimoda, G., & Kogiso, T. (2019). Effect of Serpentinite Dehydration in Subducting Slabs on

1244 Isotopic Diversity in Recycled Oceanic Crust and Its Role in Isotopic Heterogeneity of the

1245 Mantle. Geochemistry, Geophysics, Geosystems, 20(11), 5449–5472.

1246 https://doi.org/10.1029/2019GC008336

1247 Spandler, C., Hermann, J., Faure, K., Mavrogenes, J. A., & Arculus, R. J. (2008). The

1248 importance of talc and chlorite "hybrid" rocks for volatile recycling through subduction zones;

1249 evidence from the high-pressure subduction mélange of New Caledonia. Contributions to

1250 Mineralogy and Petrology, 155(2), 181-198. https://doi.org/10.1007/s00410-007-0236-2

1251 Spandler, C., Rubatto, D., & Hermann, J. (2005). Late Cretaceous-Tertiary tectonics of the

1252 southwest Pacific: Insights from U-Pb sensitive, high-resolution ion microprobe (SHRIMP)

1253 dating of eclogite facies rocks from New Caledonia. Tectonics, 24(3).

1254 https://doi.org/10.1029/2004TC001709

1255   Sun, S. -s., & McDonough, W. F. (1989). Chemical and isotopic systematics of oceanic basalts:

1256   implications for mantle composition and processes. Geological Society, London, Special

1257   Publications, 42(1), 313-345. https://doi.org/10.1144/GSL.SP.1989.042.01.19

1258   Sutherland, R., Dickens, G. R., Blum, P., & Scientists, the E. 371. (2019). Tasman Frontier

1259   Subduction Initiation and Paleogene Climate (R. Sutherland, G. R. Dickens, & P. Blum, Eds.).

1260   Proceedings of the International Ocean Discovery Program, 371: College Station, TX

1261   (International Ocean Discovery Program). https://doi.org/10.14379/iodp.proc.371.2019

1262   Taetz, S., John, T., Bröcker, M., & Spandler, C. (2016). Fluid–rock interaction and evolution of a

1263   high-pressure/low-temperature vein system in eclogite from New Caledonia: insights into

1264   intraslab fluid flow processes. Contributions to Mineralogy and Petrology, 171(11), 90.

1265   https://doi.org/10.1007/s00410-016-1295-z

1266   Taetz, S., John, T., Bröcker, M., Spandler, C., & Stracke, A. (2018). Fast intraslab fluid-flow

1267   events linked to pulses of high pore fluid pressure at the subducted plate interface. Earth and

1268   Planetary Science Letters, 482, 33–43. https://doi.org/10.1016/j.epsl.2017.10.044

1269   Tarling, M. S., Rooney, J. S., Viti, C., Smith, S. A. F., & Gordon, K. C. (2018). Distinguishing

1270   the Raman spectrum of polygonal serpentine. Journal of Raman Spectroscopy, 49(12), 1978–

1271   1984. https://doi.org/10.1002/jrs.5475

1272   Tatsumi, Y. (1986). Formation of the volcanic front in subduction zones. Geophysical Research

1273   Letters, 13(8), 717-720. https://doi.org/10.1029/GL013i008p00717

1274   Tenthorey, E., & Hermann, J. (2004). Composition of fluids during serpentinite breakdown in

1275   subduction zones: Evidence for limited boron mobility. Geology, 32(10), 865.

1276   https://doi.org/10.10.1130/G20610.1

1277   Tewksbury-Christle, C. M., Behr, W. M., & Helper, M. A. (2021). Tracking Deep Sediment

1278   Underplating in a Fossil Subduction Margin: Implications for Interface Rheology and Mass and

1279   Volatile Recycling. Geochemistry, Geophysics, Geosystems, 22(3), 1–32.

1280   https://doi.org/10.1029/2020GC009463

1281 Toffol, G., Yang, J., Pennacchioni, G., Faccenda, M., & Scambelluri, M. (2022). How to quake a

1282 subducting dry slab at intermediate depths: Inferences from numerical modelling. Earth and

1283 Planetary Science Letters, 578, 117289. https://doi.org/10.1016/j.epsl.2021.117289

1284 Ulmer, P., & Trommsdorff, V. (1995). Serpentine Stability to Mantle Depths and Subduction-

1285 Related Magmatism. Science, 268(5212), 858–861. https://doi.org/10.1126/science.268.5212.858

1286 Ulrich, M., Muñoz, M., Boulvais, P., Cathelineau, M., Cluzel, D., Guillot, S., & Picard, C.

1287 (2020). Serpentinization of New Caledonia peridotites: from depth to (sub-)surface.

1288 Contributions to Mineralogy and Petrology, 175(9), 91. https://doi.org/10.1007/s00410-020-

1289 01713-0

1290 Ulrich, M., Picard, C., Guillot, S., Chauvel, C., Cluzel, D., & Meffre, S. (2010). Multiple melting

1291 stages and refertilization as indicators for ridge to subduction formation: The New Caledonia

1292 ophiolite. Lithos, 115(1–4), 223–236. https://doi.org/10.1016/j.lithos.2009.12.011

1293 Valley, J. W., Kitchen, N., Kohn, M. J., Niendorf, C. R., & Spicuzza, M. J. (1995). UWG-2, a

1294 garnet standard for oxygen isotope ratios: Strategies for high precision and accuracy with laser

1295 heating. Geochimica et Cosmochimica Acta, 59(24), 5223-5231. https://doi.org/10.1016/0016-

1296 7037(95)00386-X

1297 van Keken, P. E., Hacker, B. R., Syracuse, E. M., & Abers, G. A. (2011). Subduction factory: 4.

1298 Depth-dependent flux of H 2 O from subducting slabs worldwide. Journal of Geophysical

1299 Research, 116(B1), B01401. https://doi.org/10.1029/2010JB007922

1300 Vitale Brovarone, A., & Agard, P. (2013). True metamorphic isograds or tectonically sliced

1301 metamorphic sequence? New high-spatial resolution petrological data for the New Caledonia

1302 case study. Contributions to Mineralogy and Petrology, 166(2), 451-469.

1303 https://doi.org/10.1007/s00410-013-0885-2

1304 Vitale Brovarone, A., Agard, P., Monié, P., Chauvet, A., & Rabaute, A. (2018). Tectonic and

1305 metamorphic architecture of the HP belt of New Caledonia. Earth-Science Reviews,

1306 178(January), 48-67. https://doi.org/10.1016/j.earscirev.2018.01.006

1307 Wada, I., Wang, K., He, J., & Hyndman, R. D. (2008). Weakening of the subduction interface

1308 and its effects on surface heat flow, slab dehydration, and mantle wedge serpentinization. Journal

1309 of Geophysical Research, 113(B4), B04402. https://doi.org/10.1029/2007JB005190

1310 Wakabayashi, J. (1992). Nappes, Tectonics of Oblique Plate Convergence, and Metamorphic

1311 Evolution Related to 140 Million Years of Continuous Subduction, Franciscan Complex,

1312 California. The Journal of Geology, 100(1), 19–40. https://doi.org/10.1086/629569

1313 Whitney, D. L., & Evans, B. W. (2010). Abbreviations for names of rock-forming

1314 minerals. American Mineralogist, 95(1), 185-187.

1315 Wu, K., Ding, X., Ling, M.-X., Sun, W., Zhang, L.-P., Hu, Y.-B., & Huang, R.-F. (2018).

1316 Origins of two types of serpentinites from the Qinling orogenic belt, central China and associated

1317 fluid/melt-rock interactions. Lithos, 302-303, 50-64. https://doi.org/10.1016/j.lithos.2017.12.019

1318 Wunder, B., & Schreyer, W. (1997). Antigorite: High-pressure stability in the system

1319 MgO☐SiO2☐H2O (MSH). Lithos, 41(1–3), 213–227. https://doi.org/10.1016/S0024-

1320 4937(97)82013-0

1321 **Figure 1.** Schematic subduction zone cross-section showing relevant locations of

1322 serpentinization and enrichments and loss of fluid-mobile elements during subduction and

1323 exhumation (after Deschamps et al., 2013; Peters et al., 2017). Large semi-transparent arrow

1324 indicates subduction and exhumation path of serpentinites, emphasizing that these rocks can

1325 encounter chemical changes on both the prograde and retrograde paths. Subduction zone after

1326 Guillot et al. (2015).

1327 **Figure 2.** (a) Regional map highlighting significant tectonic features of the SW Pacific region

1328 after Maurizot & Collot (2009) and Sutherland et al. (2019). (b) Simplified geologic map of the

1329 main island of New Caledonia showing sample locations and additional localities referenced in

1330 text. Map and units after Maurizot et al. (2020c). (c) Simplified map of Pam Peninsula showing

1331 locations of Group I serpentinites samples (this study) and prior studied samples. Unit names

1332 after Maurizot et al. (2020c) and boundaries modified after Vitale-Brovarone et al. (2018). (d)

1333 Simplified map showing sampled locations at Poadja Massif. Estimated uncertainty for

1334   Fitzherbert et al. (2004) georeferenced raster map locations (c, d) is ~500-700 m and ~200-400

1335   m, respectively. Solid black lines denote roads, solid blue lines denote streams, and dashed black

1336   line denotes high-angle normal fault.

1337   **Figure 3.** Backscatter electron images from select high-pressure serpentinites. (a) A relict

1338   chromite core surrounded by a mantle of ferritchromite and rim of chromian magnetite

1339   containing inclusions of antigorite (NC19-48). (b) Delicate veins of tremolite and antigorite cross

1340   cut a pervasively serpentinized matrix (NC19-94). (c) Skeletal chromite replaced by magnetite

1341   and antigorite (NC18-39B). (d) Inclusions of Fe-Ni alloy awaruite and antigorite in magnetite

1342   (NC18-39A). (e) Brightness-contrast enhanced image highlighting complex curvilinear

1343   retrogression of chromian magnetite to magnetite (NC19-154A). (f) Representative texture of

1344   typical replacement of olivine by antigorite and magnetite via growth from fractures (NC18-

1345   39A).

1346   **Figure 4.** Whole rock major element ratios of MgO/SiO_2 vs. Al_2O_3/SiO_2 for serpentinites and

1347   hybrid rocks in New Caledonia's HP/LT terrane. Samples are plotted with compiled data for

1348   harzburgites in the New Caledonia ophiolite (gray circles) (Liu et al., 2018; Mothersole, 2014;

1349   Mothersole et al., 2017; Ulrich et al., 2010). Also shown is the field for abyssal serpentinites

1350   from Niu (2004). Depleted mantle value from McDonough and Sun (1995), primitive mantle

1351   value from Salters and Stracke (2004). "Terrestrial array" line after Hart and Zindler (1986),

1352   Jagoutz et al. (1979), and Niu (2004).

1353   **Figure 5.** Whole rock trace and REE compositions for New Caledonia HP/LT serpentinites.

1354   Samples are plotted together (a, b) and separately into respective categories: Group I (c, d) and

1355   Group II (e, f). Trace element concentrations are normalized to primitive mantle values of Sun

1356   and McDonough (1989) (a,c,e). REE concentrations are normalized to C1 chondrite

1357   (McDonough & Sun, 1995) (b,d,f).

1358   **Figure 6.** (a) Hydrogen (δD) and oxygen ($\delta^{18}O$) stable isotope compositions of New Caledonia

1359   HP/LT serpentinites compared with existing data for serpentine in the overlying Peridotite Nappe

1360   (ophiolite groups after Cluzel et al., 2020; Ulrich et al., 2020). Weathered samples and samples

1361   for which antigorite was not specified in Cluzel et al. (2020) were excluded. Upper mantle values

1362   from Eiler (2001), Kyser and O'Neil (1984), and Mattey et al. (1994). (b) $\delta^{18}O$ versus loss on

1363   ignition (LOI) for HP/LT serpentinites. Group I serpentinites show decreasing $\delta^{18}O$ with

1364   progressive hydration whereas Group II serpentinites exhibit a narrow range of $\delta^{18}O$ values

1365   irrespective of degree of hydration. An arrow denotes the direction of $\delta^{18}O$ change when

1366   serpentine is formed at increasing temperatures in equilbirium with seawater (e.g., Saccocia et

1367   al., 2009).

1368   **Figure 7.** Representative cross-polarized photomicrographs and correlated Raman spectra for (a)

1369   two separate matrix sites in Group I sample NC18-09B and (b) one pseudomorph (red) and one

1370   vein (blue) in Group II sample NC19-152. Diagnostic Raman peaks for antigorite are highlighted

1371   tan vertical bars. All serpentine polymorphs in Group I and II were identified as antigorite

1372   (Figures S5, S6).

1373   **Figure 8.** Variations in ratios of HFSE/HREE (i.e., Yb) with increasing LOI (%) for New

1374   Caledonia HP/LT serpentinites. Variations in trace element ratios are independent of LOI value.

1375   **Figure 9.** Plots of LREE (La, Ce, Pr, and Nd) versus HFSE (e.g., Zr) for all HP/LT serpentinites,

1376   with zoomed insets in upper right of graphs to highlight trace element depleted Group II samples.

1377   Groups I and II symbols are the same as Figure 8. One sample, NC19-158 (open orange circle),

1378   has been excluded from the trendline and computed R values.

1379   **Figure 10.** Plots for investigation of FME trends among New Caledonia HP/LT serpentinites.

1380   Data are plotted against compilations for mid-ocean ridge (MOR) serpentinites (teal circles) and

1381   forearc (FA) serpentinites (tan circles) from Peters et al. (2017). (a) Ba/Yb vs. Ba, (b) Cs/Yb vs.

1382   Cs, (c) U vs. Th, and (d) Cs vs. Yb.

1383   **Figure 11.** REE compositions for New Caledonia HP/LT serpentinites. (a) Group I serpentinites

1384   plotted versus a global dataset of mid-ocean ridge serpentinites (Peters et al., 2017); (b) Group II

1385   serpentinites plotted versus harzburgite compositions from the New Caledonia ophiolite

1386   (Marchesi et al., 2009; Secchiari et al., 2020; Ulrich et al., 2010).

Figure 1.

LREE, Eu, B, Cl, U, Th, Sr

Accretionary wedge

0 km

Oceanic crust

LREE, Ti, As, Sb

Overriding plate

Lithospheric mantle

50 km

Cs, Sr, Rb, Li

Li, B

1100 °C

100 km

1000 °C

Possible source areas for serpentinites within subduction channel

Figure 2.

a (legend)

active oceanic spreading ridge · active subduction · Zealandia

land · submerged continental crust · basin with thinned continental crust · volcanic island arc

b New Caledonia HP/LT Serpentinites

Group I (n = 15)
Group II (n = 8)

Other serpentinites (this study)
Post-obduction lithologies
Peridotite Nappe
HP-LT metamorphic belt
Poya Terrane
Montagnes Blanches Nappes
Undifferentiated "basement" terranes

Kalaa-Gomen
Massif du Sud
Nouméa

Fig. 1c
Fig. 1d

50 km

c Pam Peninsula

Pacific Ocean
Pouébo Terrane
Diahot-Panié Complex
metamorphic grade
high-angle normal fault
~2 km to Balade
RPN7
Ouégoa
RPN7

Serpentinites
Group I (this study)
Spandler et al. (2008)
sample; Fitzherbert et al. (2004)
observation; Fitzherbert et al. (2004)
road

3 km

d Poadja Massif

Yambé

Serpentinites
Group II (this study)
Evans et al. (2017)
Fitzherbert et al. (2004)
approximate extent of Poadja Massif

~1 km to Diahoué

1 km

Figure 3.

Figure 4.

Figure 5.
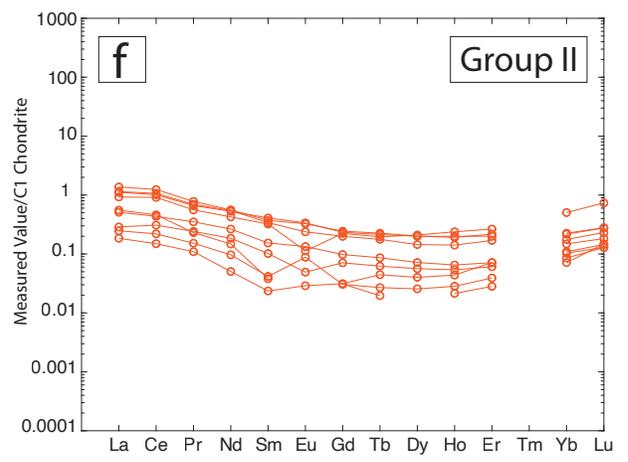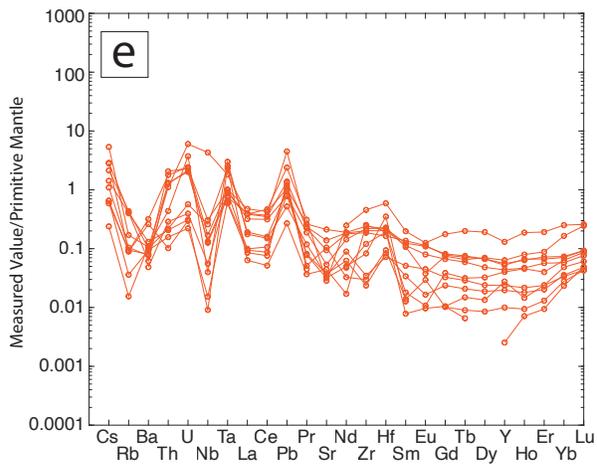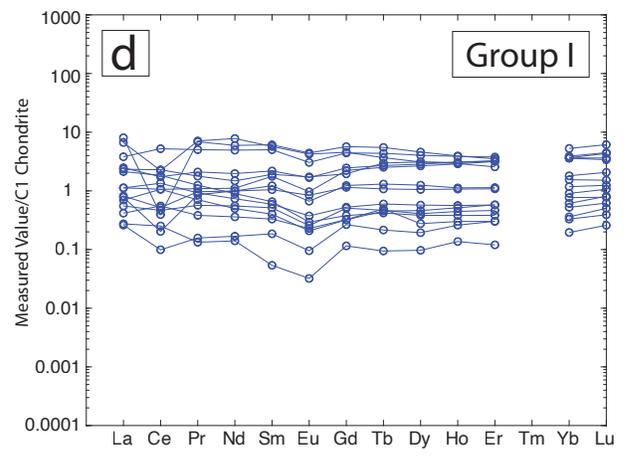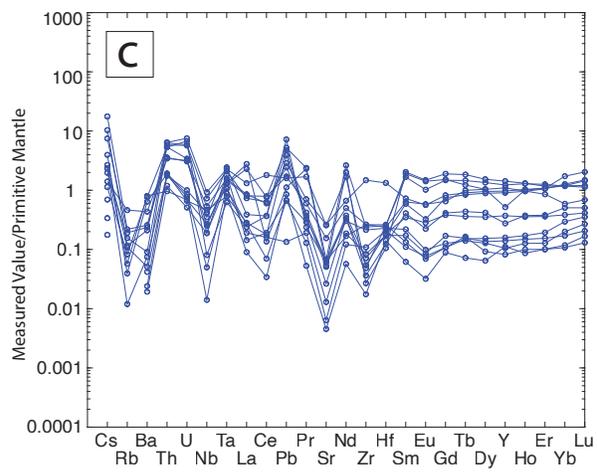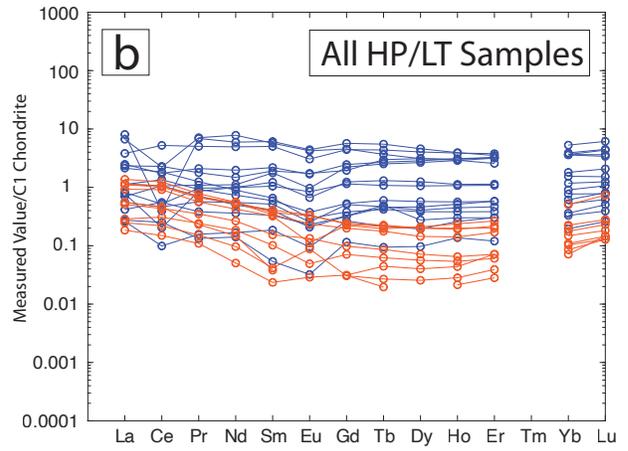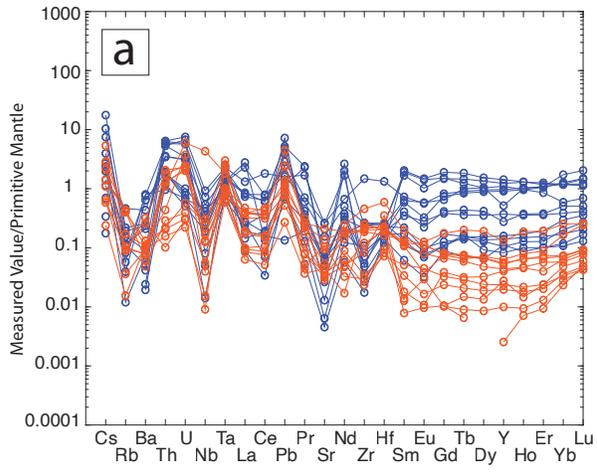
Figure 6.
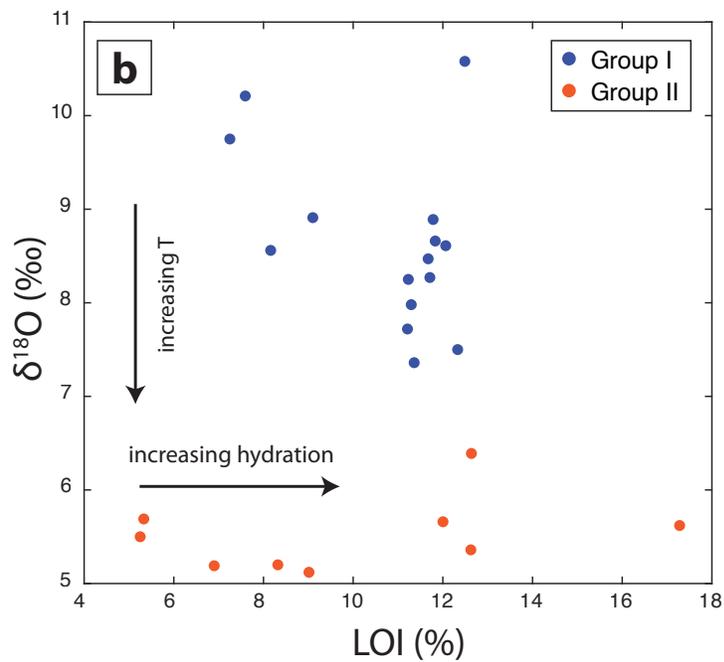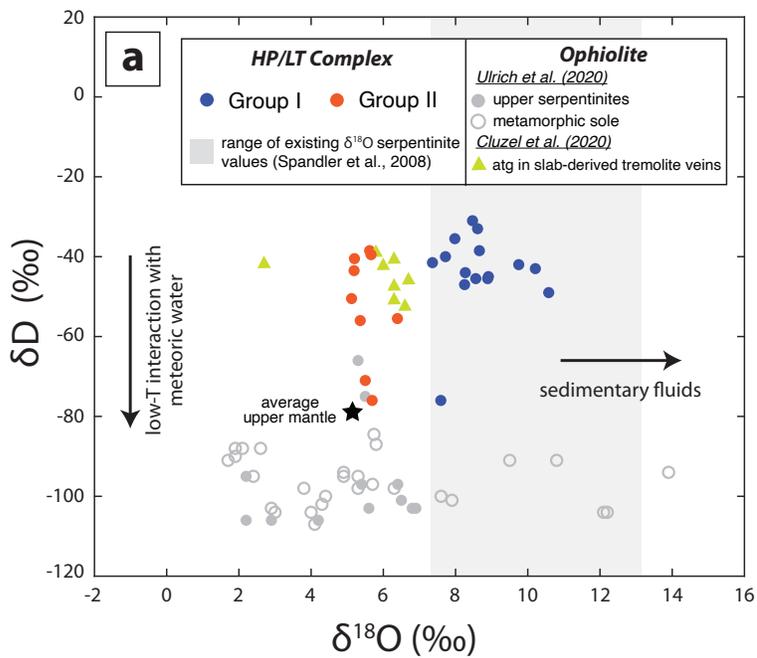
Figure 7.
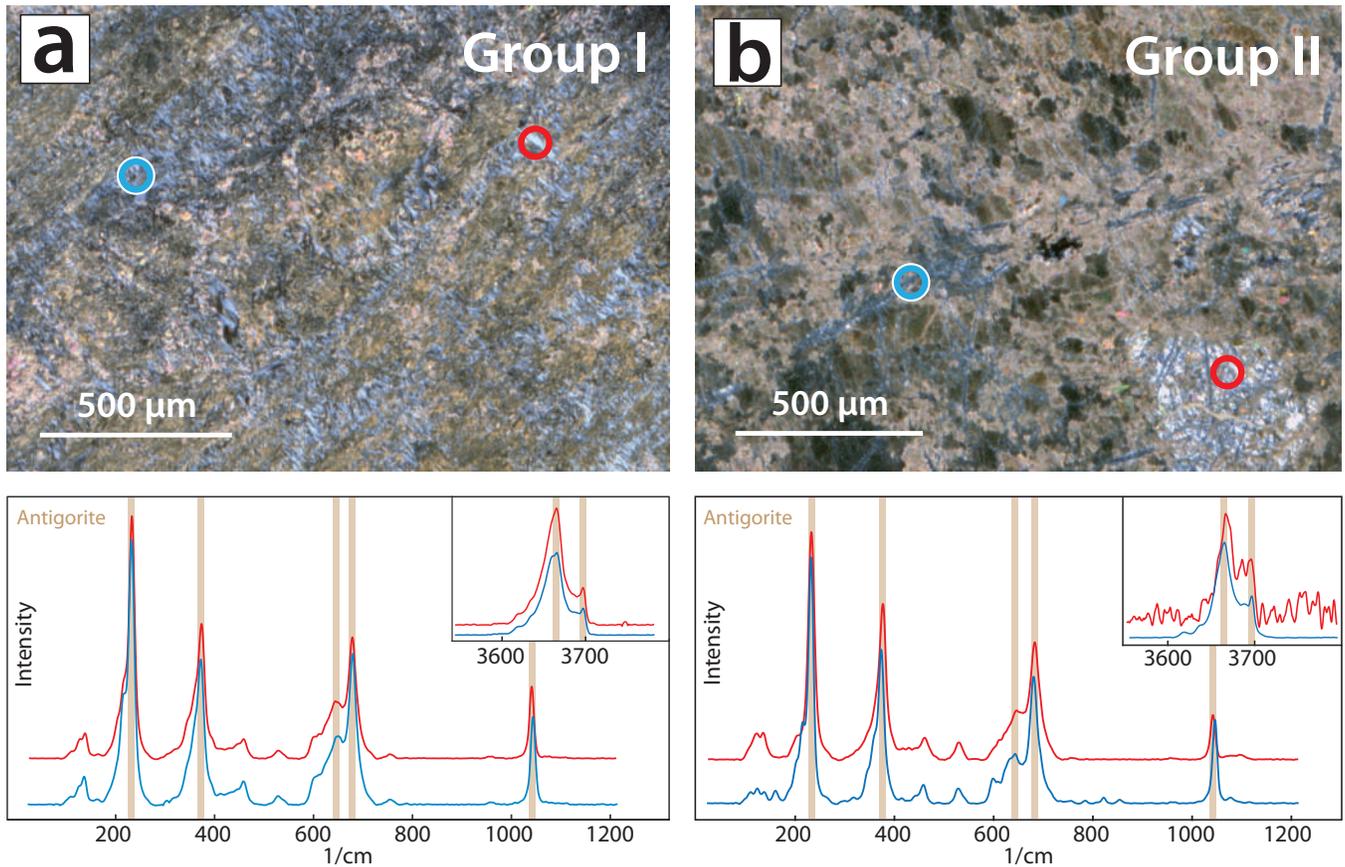
Figure 7. Representative cross-polarized photomicrographs and correlated Raman spectra for (a) two separate matrix sites in Group I sample NC18-09B and (b) one pseudomorph (red) and one vein (blue) in Group II sample NC19-152. Diagnostic Raman peaks for antigorite are highlighted tan vertical bars. All serpentine polymorphs in Group I and II were identified as antigorite (Figure S5).

Figure 8.
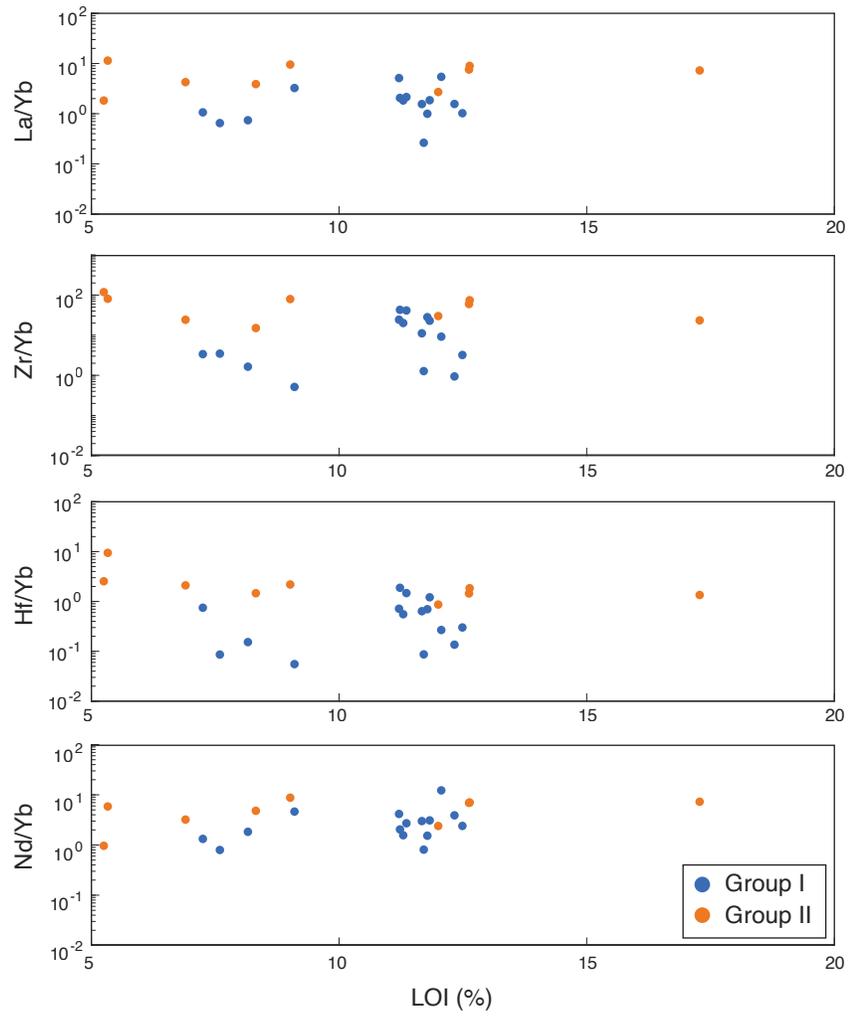
Figure 9.
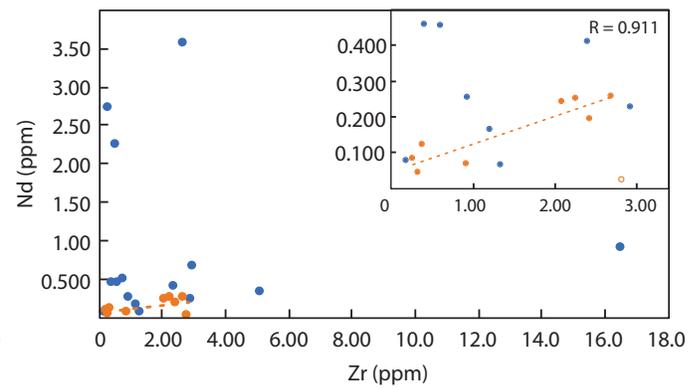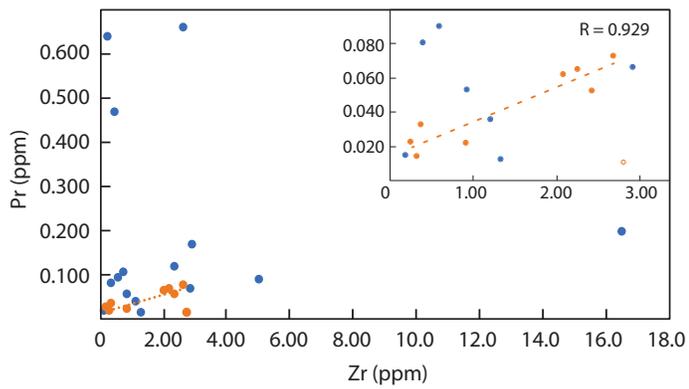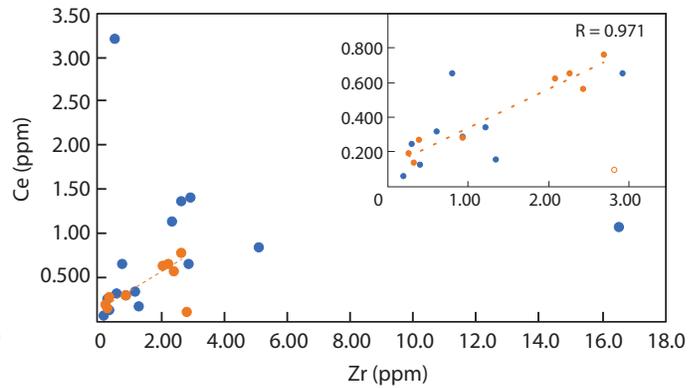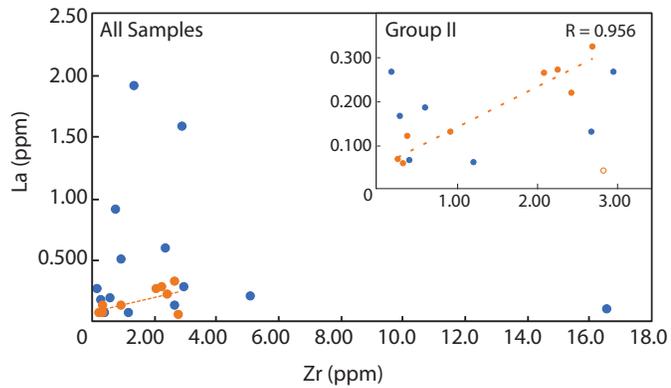
Figure 10.
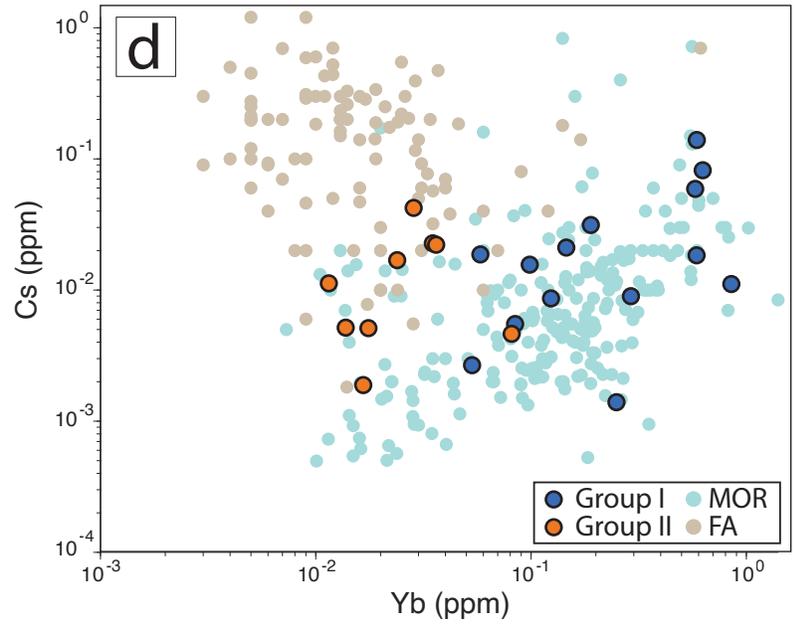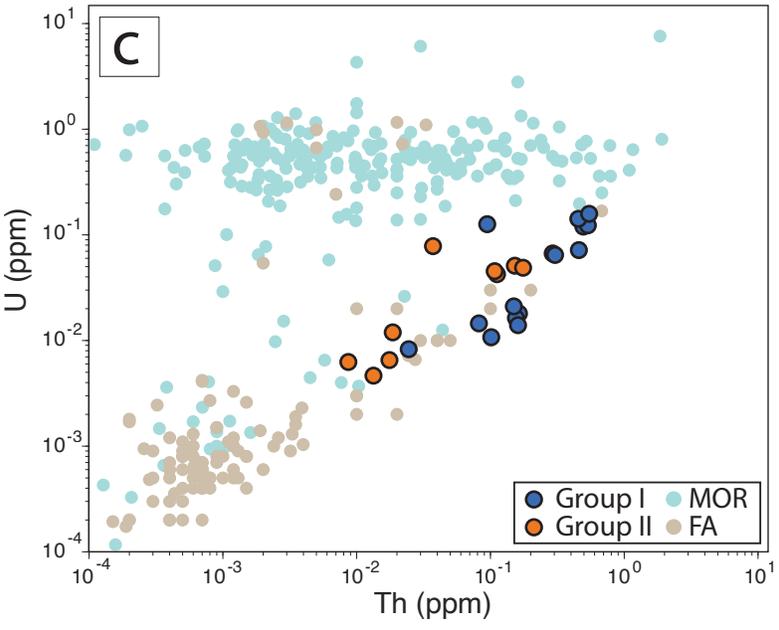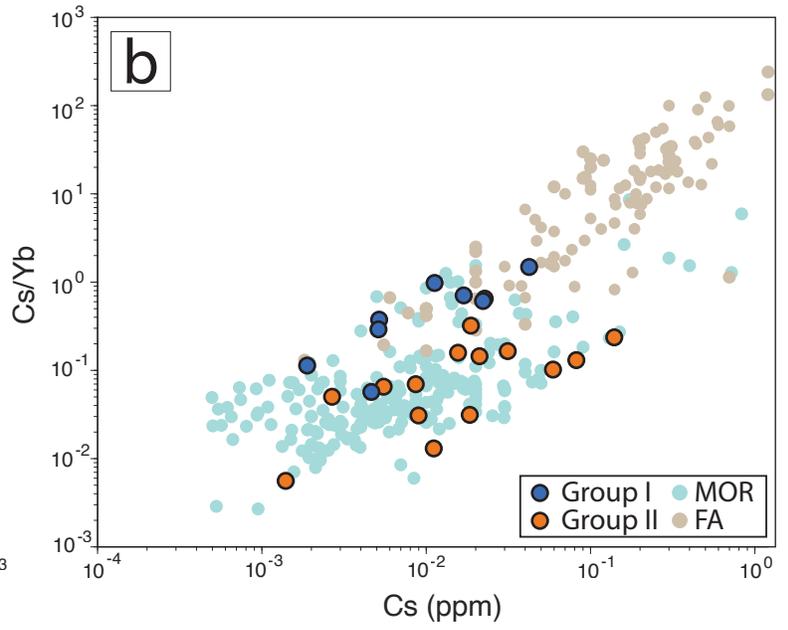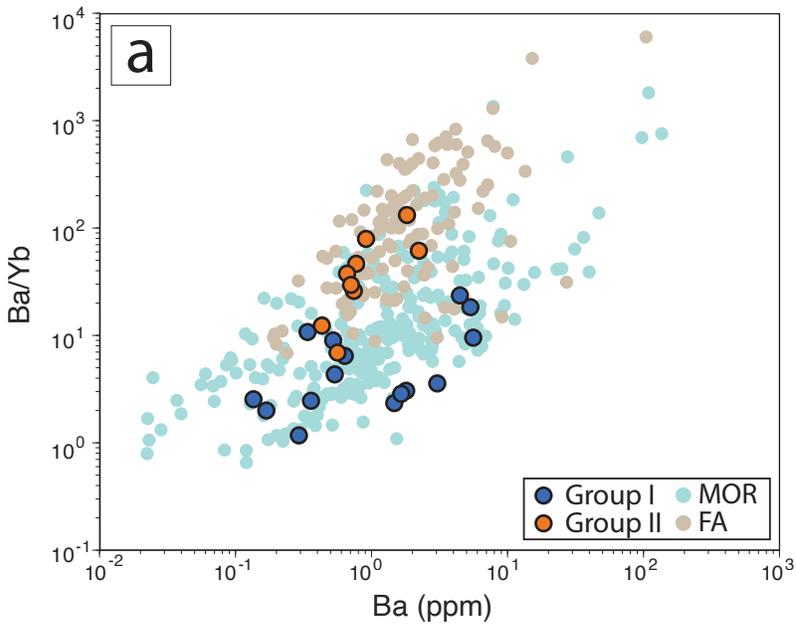
Figure 11.
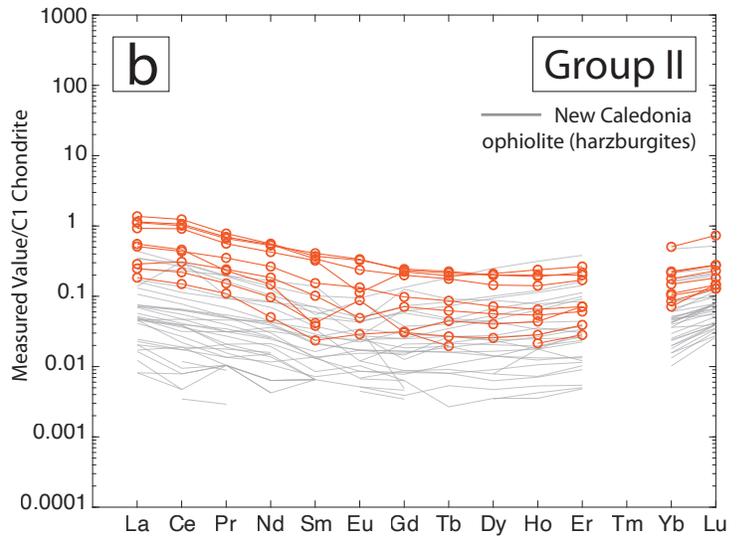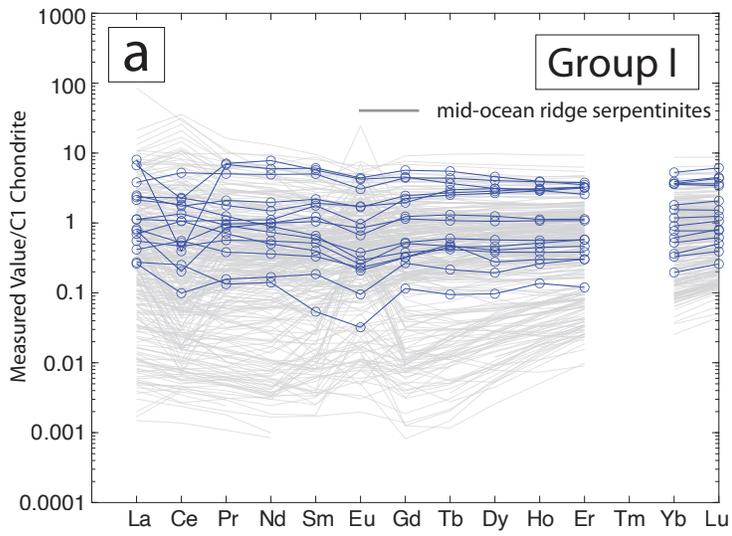
Table 1

Whole rock major and trace element compositions

Lithology	Serpentinite				
Sample name	NC18-07C	NC18-09B	NC18-15C	NC18-15D	NC18-22A
IGSN	IENHR0002	IENHR0003	IENHR0004	IENHR0005	IENHR0006
Group/Locality	I	I	I	I	I
Major elements (wt%)					
SiO_2	53.95	43.78	41.90	42.25	41.32
TiO_2	0.02	0.02	0.02	0.02	0.02
Al_2O_3	1.31	2.18	2.46	1.73	3.46
$Fe_2O_{3(T)}$	6.39	8.56	8.63	7.50	9.63
MnO	0.04	0.11	0.16	0.10	0.14
MgO	29.65	33.74	34.44	36.22	33.97
CaO	0.03	0.02	0.01	<0.01	0.01
Na_2O	0.00	0.00	0.00	<0.01	<0.01
K_2O	0.01	0.00	0.00	<0.01	<0.01
P_2O_5	0.01	0.00	0.01	<0.01	<0.01
LOI[a]	7.59	11.30	12.06	11.67	11.21
Total	98.98	99.69	99.68	99.50	99.76
Mg#	0.82	0.80	0.80	0.83	0.78
Lab	WSU	WSU	WSU	WSU	WSU
Trace elements (ppm)					
Li	n.d.	n.d.	n.d.	0.0464	n.d.
Be	n.d.	n.d.	n.d.	0.110	n.d.
Sc	7.12	10.4	8.33	7.97	11.5
V	n.d.	n.d.	n.d.	37.3	n.d.
Cr	n.d.	n.d.	n.d.	2252	n.d.
Co	n.d.	n.d.	n.d.	94.8	n.d.
Ni	n.d.	n.d.	n.d.	2384	n.d.
Cu	n.d.	n.d.	n.d.	1.89	n.d.
Zn	n.d.	n.d.	n.d.	29.4	n.d.
Ga	n.d.	n.d.	n.d.	2.73	n.d.
Rb	0.293	0.0691	0.0525	bdl	0.0992
Sr	3.27	1.34	1.43	0.274	1.29
Y	2.36	0.720	4.99	0.637	0.478
Zr	2.95	2.93	2.68	0.938	2.40
Nb	0.180	0.251	0.136	0.0356	0.519
Cs	0.0111	0.0211	0.00898	0.00551	0.0156
Ba	3.04	0.359	5.33	0.168	0.636

La	0.554	0.267	1.58	0.131	0.506
Ce	1.40	0.653	1.36	0.285	1.12
Pr	0.165	0.0655	0.659	0.0526	0.116
Nd	0.677	0.228	3.57	0.253	0.410
Sm	0.283	0.0592	0.845	0.0761	0.0970
Eu	0.0977	0.0117	0.238	0.0144	0.0165
Tb	0.108	0.0163	0.133	0.0168	0.0151
Gd	0.388	0.0638	0.906	0.101	0.0755
Dy	0.765	0.113	0.765	0.100	0.0943
Ho	0.166	0.0281	0.159	0.0244	0.0208
Er	0.515	0.0925	0.411	0.0742	0.0614
Tm	0.102	0.0169	0.0533	0.0122	0.0117
Yb	0.852	0.146	0.291	0.0843	0.0986
Lu	0.151	0.0262	0.0509	0.0152	0.0199
Hf	0.0733	0.0808	0.0779	0.0538	0.0705
Ta	0.0605	0.0998	0.0528	0.0443	0.0864
Pb	0.121	0.512	0.371	0.00955	0.173
Th	0.292	0.495	0.458	0.163	0.534
U	0.0667	0.120	0.0717	0.0180	0.123
Lab	WSU	WSU	WSU	URI	WSU

Note. n.d. = not determined; bdl = below detection limits; F&M = Franklin & Marshall College; URI =

aLOI measured by combustion

NC18-26B IENHR0008	NC18-43 IENHR000C	NC19-14 IENHR000G	NC19-48 IENHR000I	NC19-54 IENHR000J	NC19-63 IENHR000K	NC19-81 IENHR000L
38.46	46.20	42.45	54.34	42.52	42.48	51.51
0.22	0.01	0.01	0.01	0.01	0.01	<0.01
6.73	1.68	1.96	1.18	0.84	2.68	2.48
13.39	7.00	6.50	6.30	7.09	8.29	6.83
0.14	0.07	0.10	0.07	0.09	0.12	0.07
29.19	32.94	37.65	30.61	38.67	34.64	30.92
0.44	0.11	0.01	0.02	0.01	0.02	0.14
<0.01	0.18	0.03	0.01	0.03	0.02	0.02
<0.01	0.02	<0.01	<0.01	<0.01	<0.01	<0.01
<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
11.78	12.33	11.83	7.25	11.36	12.49	8.16
100.35	100.54	100.54	99.79	100.62	100.75	100.13
0.69	0.82	0.85	0.83	0.85	0.81	0.82
WSU	F&M	F&M	F&M	F&M	F&M	F&M
n.d.	5.72	0.0154	0.907	1.75	2.08	0.820
n.d.	0.846	0.0985	0.155	0.126	0.236	0.0904
13.8	10.5	6.72	4.03	6.71	11.5	7.75
n.d.	43.0	27.8	25.2	33.7	41.2	45.4
n.d.	2400	2500	2193	2861	2280	2103
n.d.	88.8	91.9	88.0	169	69.6	82.1
n.d.	2209	2299	2368	5877	1942	2288
n.d.	4.09	4.61	2.98	17.1	24.6	2.12
n.d.	39.4	35.9	40.9	106	34.0	34.7
n.d.	4.51	2.00	1.98	1.83	3.18	3.31
0.139	0.122	bdl	0.00760	0.0662	0.0250	bdl
5.54	5.40	0.135	0.0957	0.602	1.39	0.558
4.41	5.59	0.374	0.507	0.595	1.69	1.26
16.6	0.543	1.22	0.196	5.11	0.609	0.409
0.384	0.661	0.134	0.0100	3.07	0.331	0.296
0.139	0.0591	0.00267	0.0187	0.00863	0.0313	0.00140
1.80	1.65	0.135	0.522	0.537	4.47	0.292

0.587	0.903	0.0988	0.0619	0.266	0.194	0.185
1.07	3.21	0.339	0.0607	0.827	0.317	0.124
0.195	0.468	0.0355	0.0146	0.0857	0.0896	0.0798
0.902	2.26	0.165	0.0770	0.336	0.455	0.457
0.323	0.744	0.0489	0.0274	0.0878	0.155	0.181
0.0947	0.172	0.0129	0.00537	0.0211	0.0475	0.0375
0.0975	0.158	0.0177	0.00776	0.0216	0.0390	0.0472
0.489	0.881	0.0646	0.0530	0.105	0.228	0.246
0.702	0.994	0.0680	0.0477	0.141	0.260	0.308
0.169	0.211	0.0164	0.0142	0.0307	0.0586	0.0614
0.520	0.604	0.0482	0.0485	0.0921	0.173	0.182
0.0849	0.0911	0.00814	0.00799	0.0169	0.0268	0.0323
0.589	0.579	0.0532	0.0581	0.124	0.189	0.249
0.107	0.0837	0.00963	0.0124	0.0191	0.0306	0.0374
0.411	0.0786	0.0645	0.0436	0.182	0.0570	0.0382
0.0639	0.0939	0.0745	0.0362	0.0744	0.0322	0.0434
0.204	0.114	0.0482	0.0606	0.169	0.278	0.0475
0.304	0.547	0.101	0.155	0.0946	0.161	0.149
0.0646	0.158	0.0107	0.0163	0.126	0.0140	0.0210
WSU	URI	URI	URI	URI	URI	URI

= University of Rhode Island; WSU = Washington State University.

NC19-86 IENHR000N I	NC19-89 IENHR000O I	NC19-94 IENHR000P I	NC18-39A (core) IENHR0009 II	NC18-39A (rxn) IENHR0009 II	NC18-39B IENHR000A II
41.69	50.75	42.12	38.57	40.64	40.05
0.02	0.01	0.01	0.01	0.01	0.01
0.95	1.93	2.81	0.51	0.52	0.57
7.15	7.10	7.50	7.49	7.29	7.59
0.05	0.09	0.13	0.11	0.12	0.11
39.25	31.28	36.47	40.34	38.42	41.55
0.03	0.02	0.06	<0.01	0.17	0.57
0.03	<0.01	0.03	<0.01	<0.01	<0.01
<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
0.01	<0.01	<0.01	<0.01	<0.01	<0.01
11.23	9.10	11.71	12.62	12.64	9.01
100.41	100.28	100.84	99.66	99.81	99.46
0.85	0.82	0.83	0.84	0.84	0.85
F&M	F&M	F&M	WSU	WSU	WSU
0.0470	0.602	1.60	n.d.	n.d.	n.d.
0.0267	0.124	0.326	n.d.	n.d.	n.d.
8.03	10.7	9.49	6.68	7.12	7.12
30.4	45.2	58.0	n.d.	n.d.	n.d.
2454	2747	2796	n.d.	n.d.	n.d.
52.5	93.9	110	n.d.	n.d.	n.d.
1601	2386	2678	n.d.	n.d.	n.d.
3.68	11.9	0.604	n.d.	n.d.	n.d.
30.0	42.3	40.8	n.d.	n.d.	n.d.
1.33	3.64	4.44	n.d.	n.d.	n.d.
bdl	0.0364	0.0736	0.264	0.0577	0.108
bdl	1.13	1.07	1.99	4.47	2.92
0.183	6.48	4.13	0.238	0.254	0.198
1.35	0.302	0.799	2.09	2.69	2.26
0.0397	0.194	0.0574	0.0947	0.122	0.0900
bdl	0.0184	0.0819	0.0227	0.0222	0.0423
0.339	5.59	1.47	0.431	2.23	0.741

0.0649	1.91	0.165	0.265	0.325	0.271
0.154	0.242	0.652	0.620	0.760	0.654
0.0124	0.637	0.101	0.0616	0.0723	0.0646
0.0643	2.72	0.507	0.242	0.256	0.250
0.00798	0.903	0.262	0.0605	0.0550	0.0503
0.00181	0.249	0.0545	0.0190	0.0183	0.0134
0.00341	0.198	0.0909	0.00773	0.00814	0.00639
0.0230	1.13	0.444	0.0464	0.0485	0.0396
0.0239	1.13	0.655	0.0491	0.0493	0.0355
0.00747	0.215	0.158	0.0109	0.0104	0.00772
0.0192	0.563	0.507	0.0320	0.0347	0.0272
0.00424	0.0848	0.0867	0.00475	0.00558	0.00478
0.0315	0.587	0.627	0.0350	0.0362	0.0285
0.00637	0.0891	0.110	0.00679	0.00687	0.00567
0.0592	0.0325	0.0544	0.0505	0.0669	0.0624
0.0377	0.0546	0.0257	0.0356	0.0416	0.0308
0.0841	0.0793	0.340	0.0981	0.0745	0.320
0.0245	0.0820	0.454	0.112	0.152	0.108
0.00827	0.0145	0.142	0.0424	0.0510	0.0453
URI	URI	URI	WSU	WSU	WSU

NC18-39C IENHR000B II	NC19-42 IENHR000H II	NC19-152 IENHR000Q II	NC19-154A IENHR000R II	NC19-157 IENHR000S II	NC19-158 IENHR000T II	NC18-45 IENHR000D Boghen Terrane
41.61	41.96	40.80	38.86	40.64	41.98	40.26
0.01	<0.01	<0.01	<0.01	<0.01	<0.01	0.01
0.75	0.72	0.69	0.72	0.68	0.67	0.41
7.21	7.75	8.28	5.75	7.55	7.91	9.41
0.09	0.11	0.12	0.09	0.11	0.11	0.15
38.01	43.46	42.29	39.31	42.35	43.24	36.43
0.01	0.50	1.08	0.02	0.42	0.53	0.01
<0.01	0.06	0.05	0.04	0.05	0.08	<0.01
<0.01	0.01	<0.01	<0.01	<0.01	0.01	<0.01
<0.01	<0.01	<0.01	<0.01	<0.01	<0.01	<0.01
12.00	5.33	6.90	17.28	8.32	5.25	13.19
99.70	99.90	100.21	102.07	100.12	99.78	99.86
0.84	0.85	0.84	0.87	0.85	0.85	0.79
WSU	F&M	F&M	F&M	F&M	F&M	WSU
n.d.	0.717	0.481	0.0560	0.180	0.630	n.d.
n.d.	bdl	bdl	0.0160	bdl	bdl	n.d.
7.20	8.75	7.01	6.74	6.01	5.39	9.79
n.d.	21.4	12.1	17.8	19.2	17.6	n.d.
n.d.	2672	1773	2285	2117	1819	n.d.
n.d.	104	97.8	98.4	103	112	n.d.
n.d.	2331	2184	2461	2278	2555	n.d.
n.d.	10.2	2.16	2.36	3.21	2.04	n.d.
n.d.	30.1	33.2	29.3	23.3	29.7	n.d.
n.d.	0.977	1.76	0.933	0.878	0.463	n.d.
0.0598	0.254	0.0568	0.00982	0.0229	0.276	0.0814
1.13	0.670	2.20	0.724	0.788	0.934	2.19
0.289	0.0115	0.0454	0.108	0.0895	0.126	0.197
2.43	0.929	0.334	0.388	0.264	2.82	2.06
0.219	0.00648	bdl	0.0284	0.0109	0.185	0.0892
0.00464	0.0112	0.00517	0.00189	0.00512	0.0169	0.0921
0.562	0.911	1.82	0.772	0.661	0.706	3.88

0.220	0.131	0.0586	0.121	0.0682	0.0437	0.281
0.559	0.282	0.135	0.266	0.189	0.0916	0.556
0.0520	0.0214	0.0141	0.0326	0.0224	0.0102	0.0546
0.194	0.0674	0.0442	0.121	0.0844	0.0231	0.212
0.0477	0.00564	0.00616	0.0227	0.0151	0.00348	0.0367
0.00634	bdl	0.00495	0.00752	0.00277	0.00163	0.0772
0.00705	0.000708	0.000971	0.00312	0.00225	0.00161	0.00446
0.0444	0.00616	0.00608	0.0195	0.0141	0.00625	0.0273
0.0517	bdl	0.00629	0.0176	0.0139	0.00989	0.0308
0.0130	0.00117	0.00154	0.00355	0.00294	0.00239	0.00704
0.0423	0.00451	0.00627	0.0114	0.00971	0.0115	0.0230
0.00821	0.00128	0.00118	0.00187	0.00176	0.00219	0.00417
0.0813	0.0115	0.0138	0.0166	0.0176	0.0239	0.0300
0.0181	0.00363	0.00315	0.00319	0.00357	0.00449	0.00509
0.0705	0.109	0.0290	0.0224	0.0257	0.0607	0.0558
0.0403	0.105	0.0240	0.123	0.0259	0.0952	0.0311
0.0555	0.0684	0.0932	0.0369	0.0576	0.0191	0.860
0.175	0.0176	0.00867	0.0372	0.0134	0.0186	0.123
0.0488	0.00653	0.00626	0.0782	0.00465	0.0119	0.0374
WSU	URI	URI	URI	URI	URI	WSU

	Peridotite	Chlorite schist	Talc schist
NC18-46	NC18-49	NC19-178	NC19-85	NC19-169
IENHR000E	IENHR000F	IENHR000V	IENHR000M	IENHR000U
Boghen Terrane	**Serpentinite Sole**	**Kalaa-Gomen**	**I**	**I**
39.00	42.04	40.51	22.35	54.16
0.02	<0.01	<0.01	3.54	0.02
1.05	0.51	0.69	13.43	2.99
10.22	6.91	7.73	29.21	6.62
0.15	0.06	0.11	0.24	0.02
36.38	37.51	40.91	19.23	29.93
<0.01	0.04	0.66	2.47	0.03
<0.01	0.03	0.05	0.01	<0.01
<0.01	<0.01	<0.01	<0.01	<0.01
0.01	<0.01	<0.01	1.88	0.01
12.90	13.67	9.77	7.88	6.25
99.72	100.77	100.43	100.24	100.03
0.78	0.84	0.84	0.40	0.82
WSU	WSU	F&M	F&M	F&M
n.d.	1.91	1.27	7.12	0.162
n.d.	0.0109	bdl	0.252	0.0814
11.8	5.60	7.30	35.9	8.51
n.d.	19.7	26.6	448	49.0
n.d.	2314	2307	284	2336
n.d.	95.5	95.7	140	93.1
n.d.	2563	2069	247	1985
n.d.	2.55	1.82	59.7	7.40
n.d.	25.3	82.9	128	43.3
n.d.	0.645	0.879	8.47	4.28
0.144	0.00673	0.0809	0.0242	0.167
1.74	0.677	0.428	36.0	2.27
0.453	1.08	0.0220	54.8	0.327
3.12	0.497	0.353	25.0	0.256
0.155	0.00694	bdl	12.7	0.0267
0.787	0.00578	0.00336	0.00499	0.0109
3.50	0.709	0.168	12.9	0.341

0.296	0.202	0.0183	13.1	0.0523
0.600	0.170	0.0501	35.7	0.192
0.0760	0.0701	0.00444	4.29	0.0302
0.350	0.330	0.0101	20.5	0.156
0.0475	0.100	0.00109	6.42	0.0523
0.0220	0.0264	bdl	1.82	0.00663
0.00897	0.0199	0.000194	1.76	0.0118
0.0493	0.143	0.00120	9.71	0.0693
0.0652	0.110	0.000929	11.4	0.0706
0.0175	0.0273	0.000413	2.30	0.0141
0.0556	0.0752	0.00259	5.50	0.0355
0.00961	0.00983	0.000758	0.662	0.00444
0.0711	0.0544	0.00955	3.44	0.0237
0.0128	0.0109	0.00254	0.453	0.00355
0.0668	0.0437	0.0224	0.738	0.0105
0.0891	0.0525	0.0217	0.641	0.0135
0.362	0.260	0.00374	0.984	0.0827
0.0939	0.0165	0.00392	3.08	0.0223
0.0368	0.00565	0.00129	0.133	0.00457
WSU	URI	URI	URI	URI

Table 2
Oxygen and hydrogen stable isotope data.

Sample	Rock type	Group/Locality	δD$_{srp}$ (‰)	δ^{18}O$_{srp}$ (‰)
NC18-07C	serpentinite	I	-45, -41	10.2
NC18-09B	serpentinite	I	-35, -36	8.0
NC18-15C	serpentinite	I	-34, -32	8.6
NC18-15D	serpentinite	I	-33, -29	8.5, 6.7
NC18-22A	serpentinite	I	-40, -40	7.7, 6.8
NC18-26B	serpentinite	I	-44, -46	8.9
NC18-43	serpentinite	I	-75, -77	7.5
NC19-14	serpentinite	I	-38, -39	8.7, 8.8
NC19-48	serpentinite	I	-42, -42	9.8
NC19-54	serpentinite	I	-42, -41	7.4, 7.2
NC19-63	serpentinite	I	-49, -49	10.6
NC19-81	serpentinite	I	-47, -44	8.6
NC19-86	serpentinite	I	-47, -47	8.3
NC19-89	serpentinite	I	-45, -45	8.9, 8.8
NC19-94	serpentinite	I	-44, -44	8.3
NC18-39A (core)	serpentinite	II	-59, -53	5.4
NC18-39A (rxn)	serpentinite	II	-56, -55	6.4
NC18-39B	serpentinite	II	-52, -49	5.1, 5.5
NC18-39C	serpentinite	II	-39, -40	5.7
NC19-42	serpentinite	II	-73, -79	5.7
NC19-152	serpentinite	II	-43, -44	5.2
NC19-154A	serpentinite	II	-40, -37	5.6
NC19-157	serpentinite	II	-42, -39	5.2
NC19-158	serpentinite	II	-70, -72	5.5
NC19-85	chlorite schist	I		
NC19-169	talc schist	I		
NC18-45	serpentinite	Boghen Terrane	-82, -86	8.0
NC18-46	serpentinite	Boghen Terrane	-78, -79	7.0
NC18-49	serpentinite	Serpentinite Sole	-82, -82	6.5
NC19-178	serpentinite	Kalaa-Gomen Massif	-84, -85	5.7

Note. Mineral abbreviations after Whitney and Evans (2010): *Srp* serpentine, *Mag* mag

[1] >710 μm grain size fraction

[2] 125-250 μm grain size fraction

$\delta^{18}O_{mag}$ (‰)

$\delta^{18}O_{tlc}$ (‰)

9.0

1.5

2.3, 2.5
4.3[1], 4.6[2]

10.5

ɡnetite, *Tlc* talc