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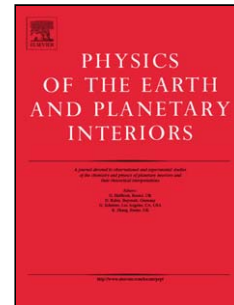
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Abstract:

Crystal / melt partition coefficients for Sr, U, and three Rare Earth Elements (La, Gd, Yb) have been measured during melting of carbonated eclogitic composition in the 4 - 7 GPa range. The experiments were conducted in a multi-anvil apparatus. Trace element concentrations were determined using electron microprobe and laser ablation ICP-MS in some cases. The obtained clinopyroxene / melt partition coefficients are very low whether the melt is silicate or carbonate in nature. This feature is due to Na presence and high vacancy concentration in the clinopyroxene. On the other hand, garnet / melt partition coefficients fall at the lower limit of or are slightly lower than literature data dealing with carbonate melt-bearing systems.

We note that clinopyroxene / carbonated melt partition coefficient values for the Rare Earth Elements cannot be modelled using the lattice strain model. The reasons for the discrepancy lie on assumptions about melt structure as well as clinopyroxene solution model. In the latter case, high Na combined with high vacancy concentration precludes ideal mixing on octahedral sites to be a valid assumption. Taking charge balance on clinopyroxene crystallographic sites does not solve the discrepancy, suggesting that melt structure can be important. Garnet / melt partition coefficients can be predicted but we note a possible influence of the melt structure. The newly obtained partition coefficients have been used to test some simple models of mantle metasomatism.

1 Trace element partitioning during partial melting of carbonated eclogites.

2

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17 **1. Introduction**

18 Although they occur rarely on the Earth's surface, carbonatitic magmas are important
19 for our understanding of deep-seated processes such as the geodynamic cycle of volatiles and
20 trace element redistribution. The nature of the carbon source as well as the petrological
21 environment where carbonatitic magma genesis takes place has been the subject of numerous
22 experimental studies focusing either on peridotitic (Wyllie and Huang, 1976 ; Eggler, 1978 ;
23 Wallace and Green, 1988 ; Falloon and Green, 1989 ; White and Wyllie, 1992 ; Dalton and
24 Presnall, 1998) or on eclogitic sources (Hammouda, 2003 ; Dasgupta et al., 2004, Yaxley and
25 Brey, 2004). All authors showed that carbonatitic magmas could be generated in the mantle at
26 P and T conditions that can vary from those of the upper mantle to those of the transition
27 zone.

28 Erupted carbonatites are often enriched in incompatible elements (e.g. Nelson et al.,
29 1988 ; Wooley and Kempe, 1989). Carbon-rich liquids are also highly mobile in the mantle
30 (Watson et al. ; 1990 ; Hammouda and Laporte, 2000), which gives them a high potential for
31 modifying lithologies they might encounter during their transport in the Earth's mantle
32 (Green and Wallace, 1988). The combination of high mobility and incompatible enrichment
33 prompted experimental studies aiming at determining trace element partition coefficients
34 between carbonatitic melts and silicate minerals (Brenan and Watson, 1991 ; Sweeney et al.,
35 1992, 1995 ; Klemme et al., 1995 ; Blundy and Dalton, 2000) in order to model trace element
36 incorporation in the melt at its source as well as element deposition during melt-rock
37 interaction. Most of the previous studies focused on minerals present in peridotites. However,
38 recent studies considered eclogitic environment as a potential source for carbonated liquids
39 (Hammouda, 2003 ; Dasgupta et al., 2004, Yaxley and Brey, 2004). In this perspective, it
40 appears that trace element partitioning data considering eclogitic mineralogy (omphacite,
41 calcic garnet, rutile) are lacking. The aim of the present study is to present experimental data

42 relevant to carbonatitic melt genesis in eclogitic environment. For this purpose, we have
43 performed partial melting experiments on trace element doped carbonated eclogitic material
44 at P and T conditions where earlier investigations showed that carbonate and / or silicate melt
45 can be produced. Run products have been analyzed by electron microprobe and laser ablation
46 ICP-MS for trace elements and partition coefficients have been determined.

47

48 **2. Experiments and analyses**

49 **2.1. Experimental protocol**

50 All experiments have been performed in multi-anvil apparatus. Starting material was a
51 trace element doped mixture of basaltic glass and natural calcium carbonate (9 : 1 in weight).
52 In order to introduce the trace elements, we mixed a synthetic basaltic glass (OTB issued from
53 the same batch as material used in Hammouda, 2003) with carbonate containing the trace
54 element mixture. Introduction of the latter was achieved by dry mixing of trace element
55 oxides to the carbonate using successive dilution steps. This procedure which was at first
56 intended for electron microprobe analysis yielded two batches of starting mixes that were
57 doped with approximately 200 and 2000 ppm La, Gd, Yb, U, Pb, Sr expressed as oxides.
58 Samples were loaded in Au₈₀Pd₂₀ capsules which were shut by arc welding.

59 High-pressure experiments were performed in the octahedral multi-anvil apparatus of
60 the Laboratoire Magmas et Volcans in Clermont-Ferrand using a Walker (1990) module and
61 following the procedure described in Hammouda (2003). High-pressure assemblies consisted
62 of Cr-doped MgO octahedra and used pyrophyllite gaskets. All experiments were performed
63 with an 18/11 assembly (octahedra side-length = 18 mm ; WC cube truncation edge-length =
64 11 mm) calibrated against the α - γ transition in the Fe₂SiO₄ system (Morishima et al., 1994)
65 and the coesite-stishovite transition (Akaogi et al., 1995). The recorded pressures of the
66 experiments are believed to be accurate within 0.5 GPa. Heating was achieved using stepped

67 LaCrO₃ tubular furnaces. The samples were placed at the center of the furnace, inside MgO
68 sleeves to prevent contact between the furnace and the gold capsules. Zirconia sleeves were
69 placed between furnaces and octahedra to minimize heat-loss. The temperature was read with
70 a W₅Re₉₅-W₂₆Re₇₄ thermocouple encased in a mullite sleeve and positioned axially in the
71 assembly, in contact with the top of the sample container. The temperature was monitored
72 using a Eurotherm 900 controller and remained constant within 1°C during the course of the
73 experiments. In some cases where the thermocouple failed, temperature was monitored using
74 the value of the electrical output power. No correction for the pressure effect on the
75 thermocouple emf was applied. Each experiment consisted of first raising the pressure to the
76 desired value before heating. Run duration was 20 hours at high temperature (Table 1). This
77 duration was considered sufficient to achieve chemical equilibrium in a system containing
78 melt and crystals at about 1300°C. For example, Blundy and Dalton (2000) used time series
79 experiments and demonstrated chemical equilibrium at 3 GPa and 1375°C in less than 20
80 hours. The runs were terminated by switching off the power of the furnace, resulting in a
81 temperature drop below 200°C in less than 2 seconds. Samples were then slowly
82 decompressed during approximately 12 hours.

83

84 **2.2. Electron probe analyses (major and trace elements)**

85 After recovery, experimental charges were mounted in resin and polished to 1/4 μm
86 grit using diamond paste. The nature of the phase present and their textural relationships were
87 determined using a scanning electron microscope at Laboratoire Magmas et Volcans.
88 Operating conditions were 15 kV accelerating voltage. Major element compositions were
89 determined with a Cameca SX100 electron microprobe at Laboratoire Magmas et Volcans.
90 Operating conditions were 15 kV accelerating voltage and beam current of 15 nA, using a
91 collapsed beam for crystalline phase and defocused beam (10 μm) for melt pools. Standards

92 used were albite (Na, Si), Al_2O_3 (Al), Fe_2O_3 (Fe), olivine (Mg, Ni), wollastonite (Ca),
93 synthetic MnTiO_3 (Mn, Ti), Cr_2O_3 (Cr) and orthoclase (K). Data reduction was performed
94 using the ZAF procedure.

95 For trace element analyses (Sr, U, Pb, La, Gd and Yb) a 25 kV accelerating voltage
96 was used with a 100 nA beam current. The three REE were chosen so as to avoid
97 interferences on the $L\alpha$ fluorescence peaks. Measurements were performed on an LLiF
98 analyzing crystal, using counting times of 100 s on peak. The three other trace elements were
99 measured on a PET crystal using the $L\alpha$ line of Sr, with counting times of 300 s and the $M\beta$
100 line of U and Pb, with counting times of 150 s. Standards were SrSO_4 for Sr, UO_2 for U,
101 home made synthetic glass for Pb, and NIST standard glasses for REE. Detection limits were
102 calculated following Ancey et al. (1987) and were less than 100 ppm for Sr and REE, 225
103 ppm for U and 350 ppm for Pb.

104

105 **2.3. Laser ablation ICP-MS analyses**

106 Trace elements in minerals and glass from experiment #273 were determined in
107 polished section using a Perkin Elmer Elan 6100 ICP-MS coupled with a femtosecond laser
108 (Pulsar 50 from Amplitude Technologies) at Laboratoire de Modélisation des Transferts
109 Géochimiques (Observatoire Midi Pyrénées, Toulouse-France, Poitrasson et al., 2005). The
110 laser was operated at 800 nm wavelength with 100 μJ /pulse energy, 3 Hz frequency and 50 fs
111 pulse duration. Crater size was 10-15 μm (Fig. 1) and ablation duration was 50-100 s.
112 Ablation was done in pure He atmosphere and the analyte was carried to the ICP torch by a
113 mixture of He +Ar. Detection limits range from 100 ppb for U to 10 ppm for Ni. NIST 610
114 glass standard (values from Norman et al. 1996) was used for calibration of relative
115 sensitivities and each analysis was normalized using CaO content for clinopyroxene, garnet
116 and melts, determined by electron microprobe. Signal intensity for major and minor elements

117 was monitored during analysis to make sure that the laser beam kept running within selected
118 mineral grain. Trace-element reductions were done with the GLITTER software (Van
119 Achterberg et al., 2001).

120

121 **3. Results**

122 **3.1 Phase relationships and major element compositions**

123 Experimental assemblages are given in Table 1 together with run conditions. Run
124 products consist of garnet, clinopyroxene, coesite (at 6 and 7 GPa) and quenched melt (Figure
125 1). Electron microprobe analyses (major elements) of run products are given in Table 2.
126 Garnets are calcium-rich whereas clinopyroxenes are sodium rich (omphacite). Additionally,
127 we found that clinopyroxenes contain a large amount of the Ca-eskola compound, a feature
128 that has been pointed out by Pertermann and Hirschmann (2002) in their study on eclogite
129 melting. Mass balance has been performed on the basis of major element compositions
130 determined by electron microprobe. Only experiments #273 and #274 could be treated. The
131 others showed strong zoning across the capsule precluding meaningful mass balance.

132 Two types of melts were found. Up to 6 GPa, we found a silicate melt with about 51
133 wt% SiO₂ at 4.5 GPa to 40 wt% SiO₂ at 6 GPa. The electron probe total ranges from about 92
134 wt% at 4.5 GPa to 82 wt% at 6 GPa. Although low total can be anticipated in glass analyses
135 using electron microprobe, the very low total found is partly attributed to dissolved CO₂ in the
136 melts. At 7 GPa, the melt is purely carbonatitic, with no silicate fraction dissolved. This melt
137 is calcium rich (Ca/ (Ca+Fe+Mg) ca. 0.77). These features are consistent with earlier work in
138 the same system (Hammouda, 2003). Melt fraction is lower at 7 GPa (ca. 10 wt%) because
139 only carbonate contributes to liquid whereas lower pressure melts (ca. 28 wt% at 6 GPa) are
140 composed of a large silicate fraction.

141

142 **3.2 Trace element concentrations and comparison between obtained partition** 143 **coefficients and literature data**

144 Table 3 presents trace element concentrations and calculated partition coefficients.
145 Lead could never be detected in the experimental run products. This absence is likely due to
146 dissolution in noble metal capsules. When available (run 273) trace element concentrations
147 obtained by the two techniques (EPMA and LA-ICPMS) are in agreement within 30% (Fig
148 2a). However, we note a systematic offset, where trace element concentrations obtained by
149 EPMA are always lower than those obtained by LA-ICPMS. In addition, it seems that heavier
150 elements display the largest offset. Important to note, however, is that calculated partition
151 coefficients determined with both techniques agree very well (Fig. 2b).

152 Calculated mineral/melt partition coefficients and comparison with literature data on
153 mineral / carbonated melt partition coefficients are presented in Figure 2a (clinopyroxene) and
154 2b (garnet). Literature data on trace element partitioning involving carbonatitic melts are
155 available on clinopyroxene (Adam and Green, 2001 ; Blundy and Dalton, 2000 ; Brenan and
156 Watson, 1991 ; Klemme et al., 1995 ; Sweeney et al., 1995, Keshav et al., 2005) and garnet
157 (Adam and Green, 2001 ; Green et al., 1992 ; Sweeney et al., 1992, 1995). Previous
158 experiments were performed between 1.5 and 6 GPa. Analytical techniques included electron
159 microprobe, ion microprobe (SIMS, Blundy and Dalton, 2000), proton probe (PIXE, Green et
160 al., 1992 ; Sweeney et al., 1992, 1995), leaching and mass balance (Brenan and Watson,
161 1991), and laser ablation ICP-MS (Adam and Green, 2001; Klemme et al., 1995; Keshav et
162 al., 2005).

163 For clinopyroxenes, we note that our values are lower by about one order of
164 magnitude when compared with literature data, except for Sr. As will be discussed later, this
165 is likely due to clinopyroxene crystal chemistry. On the other hand, our data for garnet are in

166 agreement with literature data. This general agreement for garnets is observed whatever the
167 technique used by the previous authors.

168 When possible, we also calculated clinopyroxene / garnet partition coefficients (Table
169 3). The $D_{REE}^{cpx/gt}$ have rather low values, from about 1.5 for La to 0.024 for Yb. We note that the
170 obtained values bear some resemblance to literature data on some Na- and vacancy-rich
171 clinopyroxene and Ca-rich garnet found in mantle eclogites from Roberts Victor kimberlites
172 (Caporuscio and Smyth, 1990 ; Harte and Kirkley, 1997). In particular, the higher the Na in
173 clinopyroxene the lower the REE content while Ca-rich garnet are rather rich in REE.

174

175 **4. Discussion**

176 **4.1. Mass balance and chemical equilibrium**

177 Run durations (20 hours) have been chosen to favor chemical equilibrium in the
178 samples. In order to check for the validity of our results, mass balance calculations have been
179 performed on two samples where enough data were available. In addition, those calculations
180 were used to discriminate which technique gave more consistent results when discrepancy
181 was noted. The criterion was that mass balance should reproduce the composition of the
182 starting material within 20%. Calculations performed on Sr, Gd, Yb, La and U show an
183 overall agreement between computed mass balance based on analyses of experimental
184 samples and initial concentrations in starting material (Table 4). We therefore confirm that
185 experimental samples reached equilibrium. In detail, we note that in the case of sample #273,
186 ICP-MS data are clearly overestimating true concentration for Gd and Yb, while EPMA data
187 allow to better satisfy mass balance. Strontium mass balance is satisfactory whatever the
188 analytical technique. Concerning La and U, only ICP-MS data are available. Mass balance for
189 both U and La is satisfactory.

190

191 **4.2. Lattice strain modeling and crystal chemical effects**

192 Generally, experimentally obtained trace element partition coefficients are fitted using
 193 the lattice strain model such as that developed by Wood and Blundy (1997) for REE
 194 partitioning between clinopyroxene and silicate melts. The purpose of the lattice strain model
 195 is to predict values of unknown trace element partition coefficients using major element
 196 compositions of crystals and melts equilibrated at given P and T conditions. Here we used the
 197 theoretical expressions and compared the predicted values to our experimental results. We
 198 preferred this approach, instead of the usual fitting of experimental data found in the
 199 literature, because of the small number of elements analyzed in the present investigation.
 200 Using Wood and Blundy (1997), one can predict the value of the partition coefficient D_i of an
 201 element of ionic radius r_i as a function of r_0 , the radius of the crystallographic site, E , the
 202 Young's modulus of the site, and D_0 , the strain-free partition coefficient of a cation having
 203 the radius r_0 , with

$$204 \quad D_i = D_0 \exp\left(-4\pi E N_A \left[\frac{r_0}{2} (r_i - r_0)^2 + \frac{1}{3} (r_i - r_0)^3 \right] / R T\right),$$

205 where N_A is Avogadro's number, R , the gas constant and T , the temperature in K. In the
 206 following, we discuss our results on clinopyroxene using Wood and Blundy (1997) and on
 207 garnet, using van Westrenen and Draper (2007) and Draper and van Westrenen (2007).

208

209 *Pyroxenes*

210 Following Wood and Blundy (1997) we focused on trivalent Rare Earth Elements. The
 211 predicted D_i vs. r_i relationship is shown on Figure 3a. It can be seen that the predicted
 212 partition coefficients are more than one order of magnitude larger than the actual data. This
 213 offset is larger than what is expected from Wood and Blundy (1997) model (92% probability
 214 that D_i is between 0.63 and 1.59 of the actual value). In order to gain further insight into the

215 model vs. data comparison, we have attempted to fit our REE data to the Wood and Blundy
 216 expression, keeping in mind that the fit is poorly constrained because we have only three
 217 elements. Nevertheless, La, Gd, and Yb covering almost the whole range of REE radii, we
 218 found the attempt worthwhile. The resulting fit is also shown on Figure 3. We note that the
 219 difference between the modeled and the fitted lattice strain parameters is essentially on the D_0
 220 value. The value of D_0 is defined by Wood and Blundy (1997) in the following manner

$$221 \quad D_0 = \exp\left[\frac{88750 - 65.644T + 7050P - 770P^2}{RT}\right] \cdot \frac{Mg_{melt}^\#}{X_{Mg}^{M1}}$$

222 This expression has been regressed and its coefficients (ΔH_T^0 , ΔS_T^0 , ΔV , and $\frac{\partial \Delta V}{\partial P}$) have
 223 been adjusted for volatile-free silicate melts and vacancy-free clinopyroxenes. That our D_0 is
 224 poorly predicted by the model can be due either to crystal or melt properties. To address the
 225 first possibility we used the data of Pertermann and Hirschmann (2002) on vacancy-rich
 226 clinopyroxene at 3 GPa. The effect of melt properties have been tested using the high pressure
 227 data (6 GPa) of Keshav et al. (2005) on CaO-MgO-Al₂O₃-SiO₂-CO₂ (kimberlite) system and
 228 lower pressure (2.5 GPa) data on carbonatite system by Adam and Green (2001). For all
 229 studies, we used the lattice strain model to calculate D_0 , r_0 , and E . The results are added to
 230 Figure 3. Again, we note that r_0 and E are correctly predicted while calculated D_0 is larger
 231 by a factor 1.7 to 2.1 for vacancy-rich clinopyroxene, by a factor 2.8 to 3.3 in the kimberlite
 232 melt case, and by a factor 7.6 for the carbonatite system. As in our samples, the calculated D_i
 233 values fall outside the range [0.63 – 1.59] times the true values that has been judged
 234 acceptable by Wood and Blundy (1997). Since our samples combine both characteristics (i.e.
 235 vacancy-rich pyroxenes and carbonated melts), it is easy to understand that the lattice strain
 236 model fails to predict our experimental data.

237 One reason is that the model of Wood and Blundy assumes a quasicrystalline model
 238 wherein the melt consists of pyroxene-like structural units. Rare Earth Element partitioning is

239 then described as an equilibrium between a melt and a crystal having the same stoichiometry,
 240 i.e., REEMgAlSiO₆ (Eq. 30 in Blundy and Wood, 1997). However, silica undersaturated
 241 carbonated melts such as those produced in the present investigation cannot be described in
 242 terms of pyroxene-like entities. More likely, REE solubility should be described in terms of
 243 carbonated complex because high-pressure studies on carbonate – silicate two-liquid
 244 partitioning demonstrated that REE are preferentially partitioned in carbonate melt (Hamilton
 245 et al., 1989 ; Wendtland and Harrison, 1979). Consequently, the heats, entropies, and volumes
 246 of fusion of pyroxene-like units used in fitting Wood and Blundy's equations are not
 247 applicable in our case.

248 The second reason for discrepancy between the model prediction and our data lies in
 249 the presence of vacancies in the pyroxene. The composition-activity model of Wood and
 250 Blundy (1997) uses $a_{REEMgAlSiO_6}^{cpx} = X_{REE}^{M2} \cdot X_{Mg}^{M1}$ on the assumption of a REE-bearing
 251 clinopyroxene having the stoichiometry of REEMgAlSiO₆ and ideal mixing on both M sites.
 252 However, in our case, clinopyroxene M2 site has a low mean electrostatic charge (< +1.5)
 253 because of the presence of a monovalent cation (Na) and of high vacancy concentration, while
 254 the M1 site has a high electrostatic charge (> +2.5) due to the presence of Al³⁺ associated with
 255 the vacancies (Ca-Eskola molecule, Ca_{0.5}□_{0.5}AlSi₂O₆) and not compensated by lower charge
 256 on the tetrahedra. This site filling is likely to be unfavorable for REE³⁺ incorporation in the
 257 M2 site and non-ideal mixing associated with short-range order in the clinopyroxene M sites
 258 should be anticipated. Wood and Blundy (2001) addressed the effect of electrostatic charge
 259 due to cation valence (non ideality) and they proposed a modification for the D₀ expression
 260 that takes into account the probability of having a favorable crystallographic site, in terms of
 261 charge balance. Although they were mainly interested in CaTs molecule (i.e., tetrahedral Al
 262 coupled to octahedral Al in M1 site) we attempted to model our vacancy-rich clinopyroxenes
 263 using their formulation. We found that the predicted D₀ were slightly lowered as expected for

264 clinopyroxene with little Al^{IV} but never reached our experimental values. We note from Wood
265 and Blundy (2001) that taking into account cation charge changes the D_o values for trivalent
266 cations by no more than 25% relative, either increase or decrease, depending on tetrahedral Al
267 content. Therefore, taking into account cation charge does not reconcile experimental data on
268 carbonated melt-bearing systems with predictions using lattice strain modeling. When applied
269 to Pertermann and Hirschmann's (2002) compositions, Wood and Blundy's (2001) expression
270 increases D_o values because of high Al^{IV} contents, resulting in a poorer agreement with
271 experimental values.

272

273 *Garnet*

274 Draper and van Westrenen (2007) and van Westrenen and Draper (2007) revised the
275 earlier model of van Westrenen et al. (2001) and developed two predictive models for garnet /
276 melt trivalent trace element partitioning (REE, Y, Sc). Both approaches are compared to our
277 results on REE, on Figure 3b. Overall, it seems that the models satisfactorily predict the
278 measured partition coefficients because the experimental points fall within the range defined
279 by the statistical and thermodynamic models. We note, however, that the predicted E is too
280 large resulting in impossibility of simultaneously predicting all REE D's with a single model.
281 The D_o predicted with the statistical model agrees well with the experimentally determined
282 value. Therefore, MREE and HREE can be reproduced but not LREE. On the other hand, the
283 D_o predicted with the thermodynamic model is higher by a factor of about 50 yielding in poor
284 prediction of MREE and HREE, while measured LREE falls on the predicted value. That the
285 thermodynamic model fails to predict D_o accurately suggest that either garnet or melt
286 properties are not well reproduced by the model. Concerning garnet, the critical point is the
287 presence of quadrivalent cations in the octahedral site. Si and Ti were incorporated in the
288 model for E (Young's modulus) by van Westrenen and Draper (2007) but neither the

289 thermodynamic nor the statistical model explicitly take into account the majoritic component
 290 in D_o modeling because its effect seemed to be minor. Nevertheless, our garnets are not
 291 majoritic and therefore, no compositional effect emanating from garnet is expected. On the
 292 other hand, melt composition or structure can be important. van Westrenen and Draper (2007)
 293 considered that melt effect is difficult to demonstrate given the available garnet / melt
 294 partitioning data. On the other, Draper and van Westrenen (2007) anticipate a potential effect
 295 of melt composition and structure in the presence of water. Here, we suggest that carbon
 296 presence in the melt could also have an impact on partition coefficients.

297

298 **4.3. Effect of melt composition on partition coefficients and liquid/liquid partition** 299 **coefficients**

300 Previous studies on carbonate – silicate liquid/liquid partition coefficients have shown
 301 that REE partitioning strongly depends upon pressure. At low pressure ($P < 0.1$ GPa), Veksler
 302 et al. (1998) showed that all REE but La partitioned into the silicate melt. Liquid / liquid
 303 partition coefficients, $D_{REE}^{carb/sil}$, ranged between 1.33 (La) to 0.30 (Tm). Hamilton et al. (1989)
 304 studied the effect of pressure, temperature, and melt structure on $D_{REE}^{carb/sil}$. They observed that
 305 $D_{LREE}^{carb/sil} > D_{HREE}^{carb/sil}$ at all conditions. At low pressure, REE are partitioned in the silicate melt
 306 and the trend is reversed at $P > 0.2$ GPa for LREE and $P > 0.4$ GPa for HREE. According to
 307 Hamilton et al. (1989) increasing temperature favors REE partitioning into Na-rich silicate
 308 melts but temperature has limited effect in Ca-rich system. Wendlandt and Harrison (1979)
 309 showed that REE are preferentially partitioned in carbonate melt at 1300°C and 0.5 and 2
 310 GPa.

311 The present data can be used to discuss the effect of melt composition on REE
 312 partitioning. Figure 4 shows that $D_{Gd}^{gt/liq}$ decreases smoothly with pressure for Gd in the 4.5 –
 313 7 GPa range, while $D_{Yb}^{gt/liq}$ is constant within error in the 4.5 – 6 GPa range and decreases

314 abruptly between 6 and 7 GPa. Decreasing trends for MREE and HREE have been observed
 315 by Draper et al. (2003) for garnet / melt in chondritic system. These authors attributed their
 316 observations to the onset of majoritic substitution in garnet. In the present study, we have no
 317 evidence for majoritic substitution in garnet. More likely, D_{Gd} and D_{Yb} evolution between 4.5
 318 and 6 GPa is solely due to pressure effect on garnet bulk modulus and consequently on its
 319 Young's modulus (E). Given the form of the pseudo-parabola that describe trivalent cation
 320 partition coefficients in the octahedral site of garnets (e.g. van Westrenen et al., 1999), we can
 321 anticipate that HREE will be only slightly affected by E increase because they sit on the
 322 maximum of the curve (i.e. HREE effective radii are close to the ideal value corresponding to
 323 strain-free substitution). On the other hand, as MREE sit on a descending branch, lower D_{REE}
 324 can be expected with pressure increase accompanied by higher E values.

325 The sharp decrease observed for Yb between 6 and 7 GPa (Figure 4) cannot be related
 326 solely to pressure nor to garnet composition because garnets are almost identical in the 6 and
 327 7 GPa experiments. We must therefore conclude that the change in $D_{Yb}^{gt/melt}$ is mainly due to
 328 change in liquid composition from silicate to carbonate. Wendlandt and Harrison showed that
 329 not only are the REE preferentially partitioned in carbonate melt at high pressure but also that
 330 $D_{HREE}^{carbonate/silicate} > D_{MREE}^{carbonate/silicate} > D_{LREE}^{carbonate/silicate}$. The consequence is that Yb is more retained in
 331 the melt when it becomes carbonatitic, a feature that explains the sharp decrease of $D_{Yb}^{gt/melt}$
 332 between 6 and 7 GPa.

333 Although the present study was not designed to investigate trace element liquid/liquid
 334 partition coefficients, our data allow to discuss the effect of melt nature by considering the
 335 composition of silicate melt and coexisting carbonate globules in the lower pressure run (6
 336 GPa ; 1300°C). If we assume that it is an equilibrium texture, we can calculate the following
 337 partition coefficients using electron probe analyses : $D_{Sr}^{carb/sil} = 1.58$; $D_U^{carb/sil} = 1.38$;
 338 $D_{La}^{carb/sil} = 1.96$; $D_{Gd}^{carb/sil} = 1.83$; $D_{Yb}^{carb/sil} = 1.69$. We find that Sr is preferentially partitioned into

339 carbonate melt in agreement with previous investigators (Veksler et al., 1998). In the case of
340 REE, we agree with Wendlandt and Harrison (1979) and Hamilton et al. (1989) in that REE
341 are preferentially partitioned in carbonate melt at high pressure. Therefore, this trend appears
342 to be robust at mantle pressure, whatever the melt composition. Almost all previous studies
343 used alkali-rich melt compositions, mainly because the authors were interested in direct
344 comparison with natural melt compositions. In our case, melts are poorer in alkalis, mainly
345 because of the presence of Na-rich clinopyroxene in the experimental assemblages.
346 Nevertheless, high-pressures favor REE partitioning into the carbonate melt.

347

348 **5. Implications for mantle metasomatism**

349 Melts produced by carbonated eclogite melting might be liberated from the slab and
350 interact with surrounding mantle. The processes involved are illustrated in Figure 5 on the
351 basis of REE, Sr and U only. (We are presently lacking information on other elements.) In our
352 models, we considered a starting composition of 90 percent basalt + 10 percent carbonated
353 pelagic sediments (in weight). Initial REE, Sr and U concentrations were taken from
354 McDonough and Sun (1989) and Planck and Langmuir (1998) for basalt and sediments,
355 respectively. The trace element spectrum of the starting mixture is shown on Figure 5a. All
356 simulations considered batch-melting hypothesis.

357 The spectra of melts produced at 6 and 7 GPa, are almost identical (Fig. 5a). The
358 contrast in major element composition between melts produced at 6 GPa (silicate) and 7 GPa
359 (carbonate) appears to have limited impact. Main features are enrichment of 80 – 100 for the
360 most incompatible elements, and enrichment of slightly less than 10 for HREE, relative to
361 primitive mantle (McDonough and Sun, 1995). HREE fractionation relative to LREE is due to
362 garnet presence in the residue. Positive U and Sr anomalies reflect the simulated source
363 characteristics.

364 Next we consider two end-member models for a carbonatite melt interacting with
365 peridotite mantle (Figure 5b). In the first case, a low melt fraction (1 wt%) is simply added to
366 the peridotite. The resulting REE spectra exhibit slight LREE enrichment relative to HREE
367 ($[La/Yb]_N < 2$). In the second type of model, we considered large volumes of carbonatitic
368 melt percolating through the mantle and subsequent melt-mineral equilibrium. In this model
369 we used the partition coefficient fitted from the data of Adam and Green (2001) for garnet,
370 and from Blundy and Dalton (2000) for clinopyroxene. The percolated mantle contains 15
371 wt% clinopyroxene and 8 wt% garnet. The calculated REE spectrum exhibits HREE
372 enrichment relative to LREE ($[La/Yb]_N < 0.01$) with $[LREE]_N < 1$. Very low $[LREE]_N$ is due
373 to low values of the clinopyroxene / melt and garnet / melt partition coefficients for the LREE
374 and the assumption of very large volume of percolating melt that leached the mantle. A more
375 plausible scenario is that of infinite mantle reservoir. In that case, the LREE part of the
376 spectrum levels off with values slightly less than 1.

377 The third step of the model consists in remelting the mantle region metasomatized by
378 the carbonatitic melt (Figure 5c). One case considers mantle wherein melt has been simply
379 added (melting degree 0.5 and 1.5 wt%), whereas the other case (1.5 wt% melting degree) is
380 for the mantle region equilibrated with the percolating melt. The first model yields a steep
381 pattern with high LREE enrichment relative to HREE ($[La/Yb]_N$ about 40 to > 100). On the
382 opposite, the second model yields pattern with $[REE]_N$ in the range 5 – 10 exhibiting slight
383 MREE enrichment. This type of spectrum is unknown for natural melts but some alkaline
384 cumulates have been found with similar patterns (Grégoire et al., 1998).

385 The REE pattern obtained by melting of mantle with simple addition of 1 wt%
386 carbonated melt has about the same level of HREE enrichment and twice LREE levels when
387 compared to melting of fertile pyrolitic mantle (with 1.5 wt% melting, Fig. 5c). Our modeled
388 melts bears strong resemblance with those of kimberlite particularly when very small melting

389 degrees are considered. Modeling of trace element patterns of kimberlites has been attempted
390 by Keshav et al. (2005). Those authors modeled melting of a refractory mantle portion
391 mesomatized by silicate melt by 1 wt% melting and found that similarities with kimberlites
392 could be obtained only after repeated steps of melting and metasomatism. Here we find that a
393 single metasomatic stage is sufficient in order to obtain kimberlite REE signatures.

394

395 **6. Conclusions**

396 We have obtained new partition coefficients values relevant to melting of subducted
397 carbonate-bearing eclogitic composition in the 4 – 7 GPa range. The nature of Na- and
398 vacancy-rich clinopyroxenes yields very low values for crystal / melt partition coefficients of
399 REE, U, and Sr to a lesser extent, compared to literature data. Garnet / melt partition
400 coefficients fall at the lower limit of literature data dealing with carbonate melt-bearing
401 systems.

402 Clinopyroxene / carbonated melt partition coefficient values for the Rare Earth
403 Element cannot be modeled using the lattice strain model (Wood and Blundy, 1997) in its
404 current state, mainly because the pyroxene and melt respective solution models differ strongly
405 in the present case. In addition, ideal solution on each octahedral (M) site in Na- and vacancy-
406 rich clinopyroxenes might no longer be a valid assumption. Taking into account cation charge
407 budget on octahedral sites (Wood and Blundy, 2001) does not allow reducing the discrepancy
408 significantly. We also find that garnet partition coefficients are not predicted in a satisfactory
409 manner, suggesting that carbonated melts have an effect that is not included in the models in
410 their present state as suspected in the case of water by Draper and van Westrenen (2007).

411 Simple modeling of single stage reaction of melts liberated by carbonated eclogites in
412 the mantle shows that it possible to create sources whose subsequent melting might yield
413 liquid bearing strong resemblance with kimberlites, on the basis of REE pattern, mainly.

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Table 1

Details on experimental run conditions and resulting phase assemblages and phase proportions.

Run#	P (GPa)	T (°C)	t (h)	Result	gt	cpx	coes	melt	Σ sq. res.
258	4.5	1300	20	gt-cpx-sil melt					
259	5.0	1300	20	gt-cpx-sil melt					
273	6.0	1300	20	gt-cpx-coes-sil melt	23.76	44.58	2.23	27.78	1.415
274	7.0	1300	20	gt-cpx-coes-carb melt	35.06	47.13	6.62	10.25	0.306

Table 2

Major element composition of selected phases in experimental assemblages. (n.a., not analyzed.)

Oxide (wt %)	SiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	Total
ME258 (4.5 Gpa - 1300°C)								
CPX	51.1081	16.1666	8.882	6.5218	14.1514	3.4629	n.a.	100.3045
GARNET	39.822	22.2105	18.1422	8.1454	12.579	0.0383	0.0151	100.9525
MELT	51.2436	15.0045	7.6117	3.1035	13.0825	2.2664	0.1731	92.4854
ME259 (5 Gpa- 1300°C)								
CPX	50.8388	15.333	7.732	6.9036	15.4316	3.5306	n.a.	99.7696
GARNET	39.317	22.0813	18.2679	8.161	12.0877	0.0688	0.0242	100.0079
MELT	47.9129	13.8499	8.9488	3.0546	14.5676	2.0411	0.1983	90.5732
ME273 (6 Gpa - 1300°C)								
CPX	52.716	17.598	5.565	6.5	13.27	4.196	n.a.	99.857
GARNET	39.168	22.195	15.111	8.731	13.62	0.109	n.a.	98.938
MELT	39.78	10.193	9.324	3.127	18.247	1.658	0.118	82.45
ME274 (7 Gpa - 1300°C)								
CPX	54.494	15.607	5.981	6.702	12.651	5.49	n.a.	100.929
GARNET	40.065	21.81	15.282	9.017	13.709	0.125	0.035	100.046
MELT	0.086	0.161	7.236	4.555	40.67	0.472	0.066	53.246

Table 3

Structural formulae and site filling for clinopyroxenes and garnets from Table 2. All iron was assumed to be in the form Fe²⁺. For clinopyroxene, saturated M1 octahedral site was considered and all vacancies were assigned to M2 site.

CPX	ME258	ME259	ME273	ME274	GARNET	ME258	ME259	ME273	ME274
P (GPa)	4.5	5	6	7	P (GPa)	4.5	5	6	7
T (°C)	1300	1300	1300	1300	T (°C)	1300	1300	1300	1300
Si	1.842	1.843	1.870	1.918	Si	2.991	2.983	2.976	3.010
AlIV	0.158	0.157	0.130	0.082	Al	1.967	1.975	1.988	1.931
M1					Fe	1.140	1.159	0.960	0.960
AlVI	0.529	0.498	0.605	0.566	Mg	0.912	0.923	0.989	1.009
Mg	0.350	0.373	0.344	0.352	Ca	1.012	0.983	1.109	1.103
Fe	0.121	0.129	0.051	0.083	Na	0.006	0.010	0.016	0.018
M2					K	0.001	0.002	0.000	0.003
Ca	0.547	0.599	0.504	0.477	sum cations	8.029	8.036	8.038	8.035
Na	0.242	0.248	0.289	0.375	Ca%	33.04	32.07	36.26	35.91
Fe	0.147	0.106	0.114	0.093	Mg%	29.76	30.11	32.33	32.85
Mg	0.000	0.000	0.000	0.000	Fe%	37.20	37.82	31.40	31.24
sum cations	3.936	3.953	3.907	3.945	Mg#	44.45	44.32	50.73	51.25
vac	0.064	0.047	0.093	0.055					
Mg#	56.68	61.41	67.55	66.63					

Table 4

Trace element concentrations (in ppm) and resulting partition coefficients. Number in brackets indicate the uncertainty (1 sigma) on the last reported digits; n is the number of analyses in each phase.

By electron microprobe**Concentrations**

run#	ME258			ME259		
P (GPa)	4.5			5		
T (°C)	1300			1300		
phase	silicate melt	Garnet	D gt/silicate melt	silicate melt	Garnet	D gt/silicate melt
n	4	7		4	4	
Sr	728(47)			700(61)		
U	1023(83)			650(63)		
La	690(47)			778(29)		
Gd	330(33)	234(64)	0.71(27)	300(22)	170(51)	0.57(21)
Yb	190(83)	679(219)	3.57(271)	125(35)	458(45)	3.66(1.39)

run#	ME273							
P (GPa)	6							
T (°C)	1300							
phase	silicate melt	carbo. melt	Pyroxene	Garnet	D cpx/silicate melt	D cpx/carbo. melt	D gt/silicate melt	D gt/carbo. melt
n	11	2	5	6				
Sr	5911(44)	9385(2595)	224(21)		0.038(4)	0.024(9)		
U	3680(76)	5080(2036)						
La	4738(30)	9265(2680)						
Gd	5012(53)	9950(3041)	166(8)	1568(145)	0.033(2)	0.017(6)	0.31(3)	0.16(6)
Yb	1702(27)	2985(587)	148(24)	6135(121)	0.087(15)	0.050(18)	3.60(13)	2.06(44)

run#	ME274				
P (GPa)	7				
T (°C)	1300				
phase	carbo. melt	Pyroxene	Garnet	cpx/carbo. melt	D gt/carbo. melt
n	11	5	2		
Sr	18845(2530)	422(50)		0.022(6)	
U	12713(2813)				
La	16408(2682)				
Gd	11643(1790)	142(47)	1670(293)	0.012(6)	0.14(5)
Yb	1855(253)	106(41)	3877(450)	0.057(30)	2.09(53)

By laser ablation ICP-MS - sample ME273**Concentrations**

n	silicate melt	pyroxene	garnet
	3	3	2
Sr	6320(152)	137(10)	15.3(27)
U	5475(252)	1.7	18.7(2)
La	5221(44)	9.1(1)	6(1)
Gd	6361(180)	189(1)	2033(55)
Yb	2459(43)	205(12)	8355(134)

Partition coefficients - sample ME273

D cpx/melt	D gt/melt	D cpx/gt
0.022(2)	0.0024(5)	9.0(2.2)
0.00031(2)	0.0034(2)	0.091(1)
0.00175(3)	0.0012(3)	1.51(33)
0.030(1)	0.32(2)	0.093(3)
0.083(6)	3.4(1)	0.024(2)

Table 5
Mass balance calculations on experimental samples.

sample #273 : 6 GPa - 1300°C									
		EPMA	EPMA	EPMA	ICPMS	ICPMS	ICPMS	ICPMS	ICPMS
	fractions	Sr	Gd	Yb	Sr	La	Gd	Yb	U
Coesite	2.23	0	0	0	0	0	0	0	0
Clinopyroxene	44.58	224	166	148	137	9	189	204	1.7
Garnet	23.76	0	1568	6135	15	6	2033	8355	19
Melt	27.78	5911	5012	1702	6320	5221	6361	2459	5475
resulting balance		1742	1839	1996	1820	1456	2334	2759	1526
starting concentration		1757	1768	1740	1757	1739	1768	1740	1765
offset (rel%)		-0.86	4.01	14.74	3.61	-16.29	32.02	58.58	-13.53
sample #274 : 7 GPa - 1300°C									
		EPMA	EPMA	EPMA					
	fractions	Sr	Gd	Yb					
Coesite	6.62	0	0	0					
Clinopyroxene	47.13	422	142	106					
Garnet	35.06	0	1670	3877					
Melt	10.25	18845	11643	1855					
resulting balance		2131	1846	1599					
starting concentration		1757	1768	1740					
offset (rel%)		21.26	4.40	-8.08					

Table 6

Parameters obtained using the lattice strain model for trivalent cations in clinopyroxene and garnet. Top : fit on the experimental data on La, Gd, Yb. Bottom : calculated parameters using predictive model of Wood and Blundy (1997) for cpx and van Westrenen and Draper (2007, D_o thermodynamical) and Draper and van Westrenen (2007, D_o statistical) for garnet.

Exp #	P (GPa)	T (°C)	CPX			GARNET			
			r_o (Å)	D_o	E^{3+} (GPa)	r_o (Å)	D_o exp.	E^{3+} (GPa)	
Experimental									
273	6	1300	0.945	0.0871	251	0.945	3.46		572
Model									
273	6	1300	0.974	1.46	303	0.922	D_o thermo. 175	D_o stat. 7.69	664
274	7	1300	0.977	1.47	310	0.917	147	6.83	664

Figure captions

Figure 1 : **(a)** Backscattered electron micrograph of run products showing coesite needles (darkest gray), Na-rich clinopyroxene (dark gray), garnet (lightest gray) in glassy matrix. **(b)** Close-up view of the region shown in (a) illustrating the size of the craters due to laser ablation.

Figure 2 : Obtained partition coefficients and comparison with literature data (references used are given in the text).

Figure 3 : Comparison between experimentally obtained partition coefficient values and values predicted using lattice strain modeling. Ionic radii are those of Shannon (1976). **(3a)** *Clinopyroxenes* : experimental data (exp.) are compared to predictions (pred.) using Wood and Blundy (1997). In addition to data obtained in the present investigation, references used are AG, Adam and Green (2001) ; K, Keshav et al. (2005) ; PH, Pertermann and Hirschmann (2002). **(3b)** *Garnets* : experimental data (exp.) are compared to predictions using the statistical approach (pred. stat) of Draper and van Westrenen (2007) and the thermodynamic approach (pred. thermo) of van Westrenen and Draper (2007). See text for further discussion.

Figure 4 : Partition coefficient values for Gd and Yb in garnet as a function of pressure.

Figure 5 : Modeling of REE spectra associated to mantle metasomatism due to melting of MORB + carbonate mixtures. All spectra are normalized relative to primitive mantle (McDonough and Sun, 1995). **(5a)** Modeled REE spectra of a source composed of 0.9 MORB + 0.1 carbonated sediments as used in the experiments and liquids produced by 10 wt%

melting degree (batch melting) at 6 and 7 GPa. **(5b)** Modeled REE spectra of a lherzolitic mantle after interaction with carbonatitic melts produced at 7 GPa. Two cases have been considered: simple adjunction of 1 wt% melt (10 wt% melting degree) to mantle; melt mantle equilibration wherein the mantle portion is impregnated by an interconnected melt network (also produced by 10 wt% melting degree). In the latter case, partition coefficients used were those of Blundy and Dalton (2000) for clinopyroxene and Adam and Green (2001) for garnet. Both melt types are used in the models detailed in the next panel. **(5c)** Modeled REE spectra for melting of a metasomatized mantle portion and comparison with low degree melting of fertile pyrolite. The following cases are illustrated: (■) 1.5 wt% melting of pyrolite with 1 wt% carbonatite melt added; (▲) 0.5 wt% melting of pyrolite with 1 wt% carbonatite melt added; (○) 1.5 wt% melting of pyrolite equilibrated with interconnected carbonatitic melt network treated as an infinite reservoir; (Δ) 1.5 wt% melting of pyrolite. All models assume batch melting.

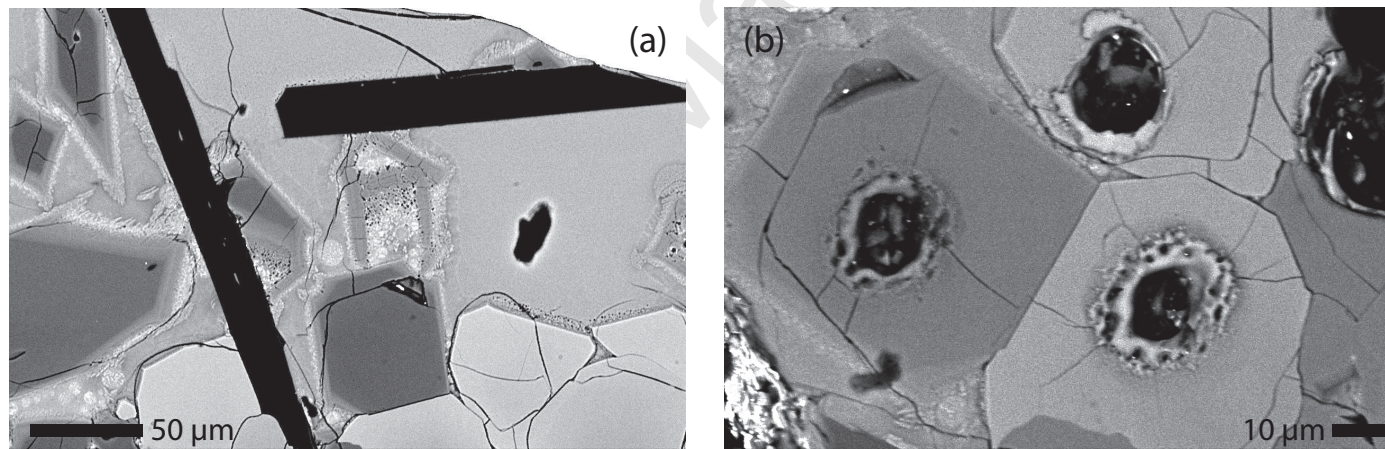


FIGURE 1

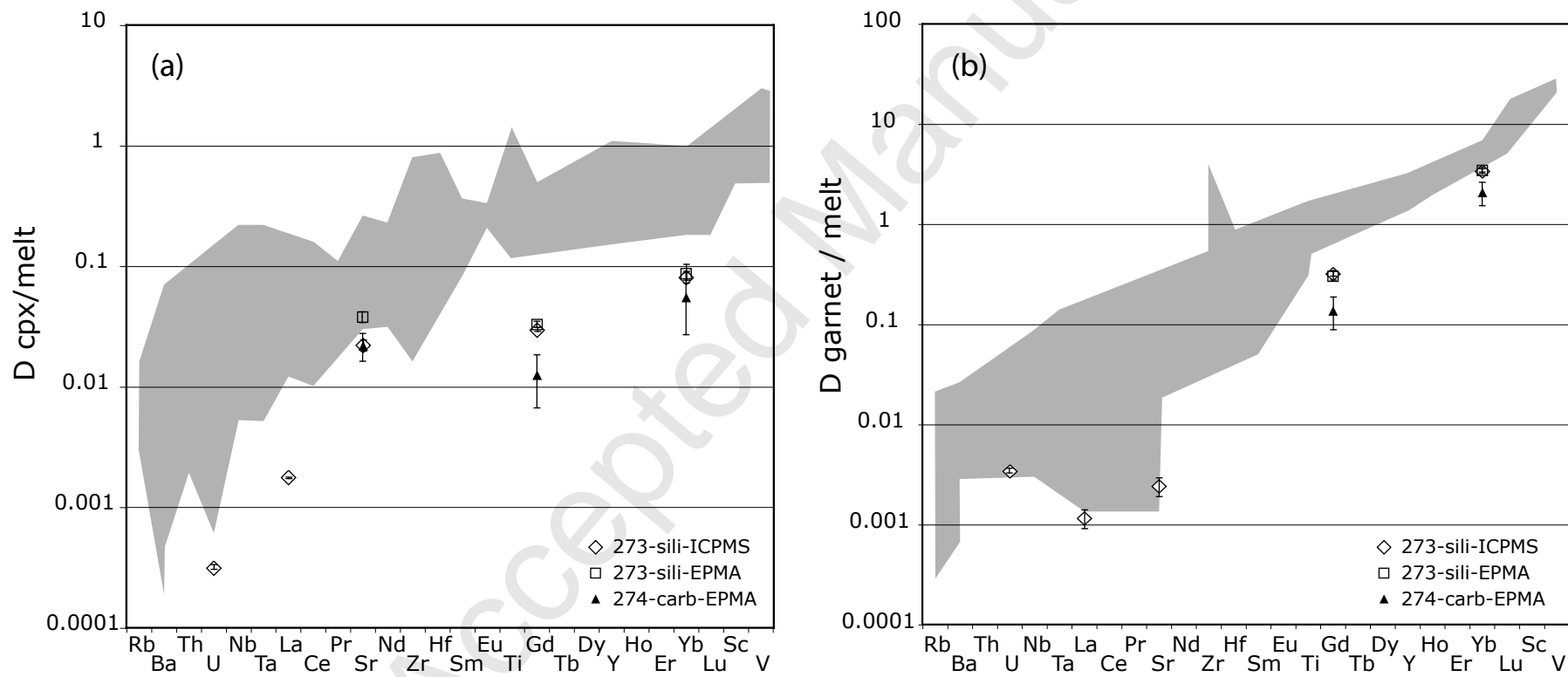


FIGURE 2

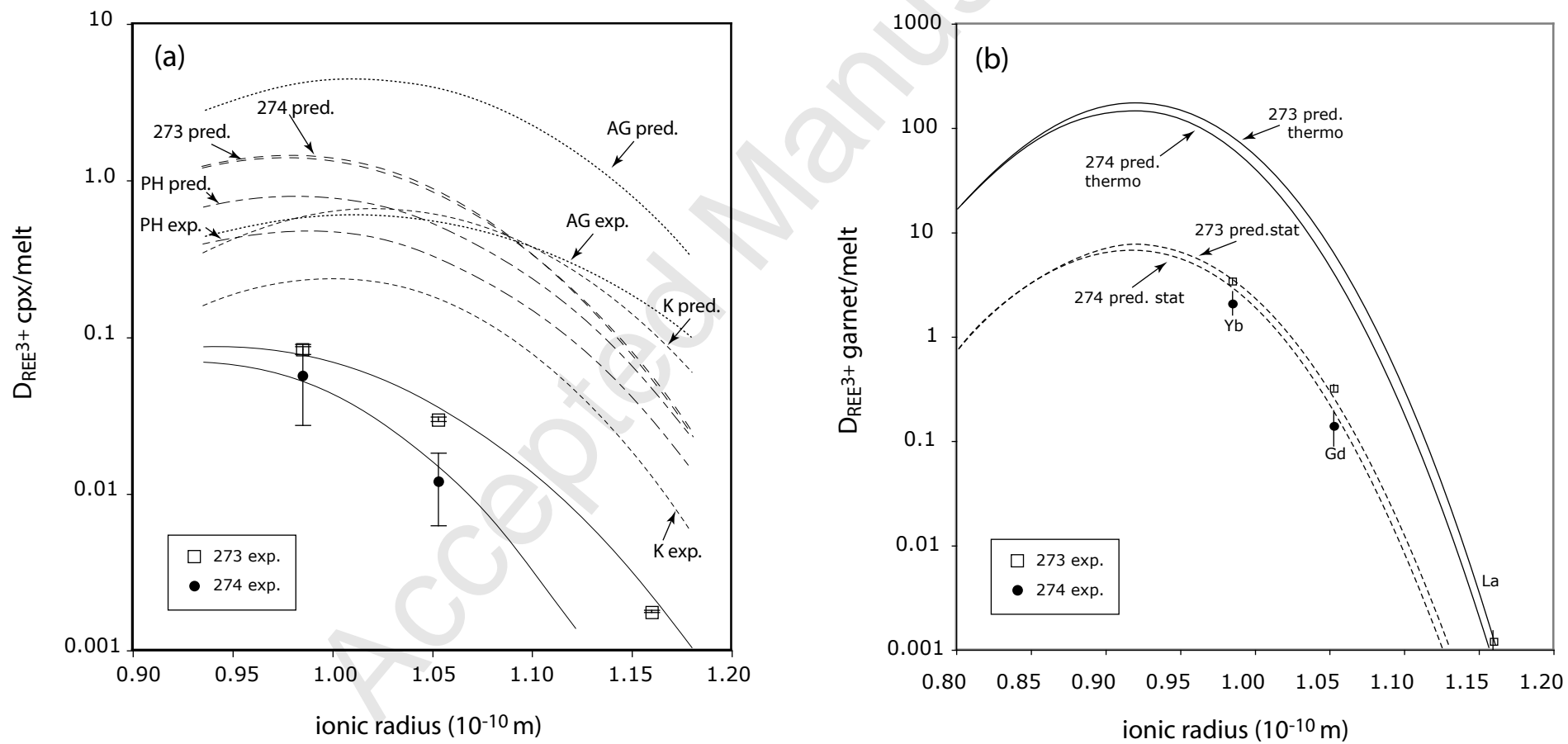


FIGURE 3

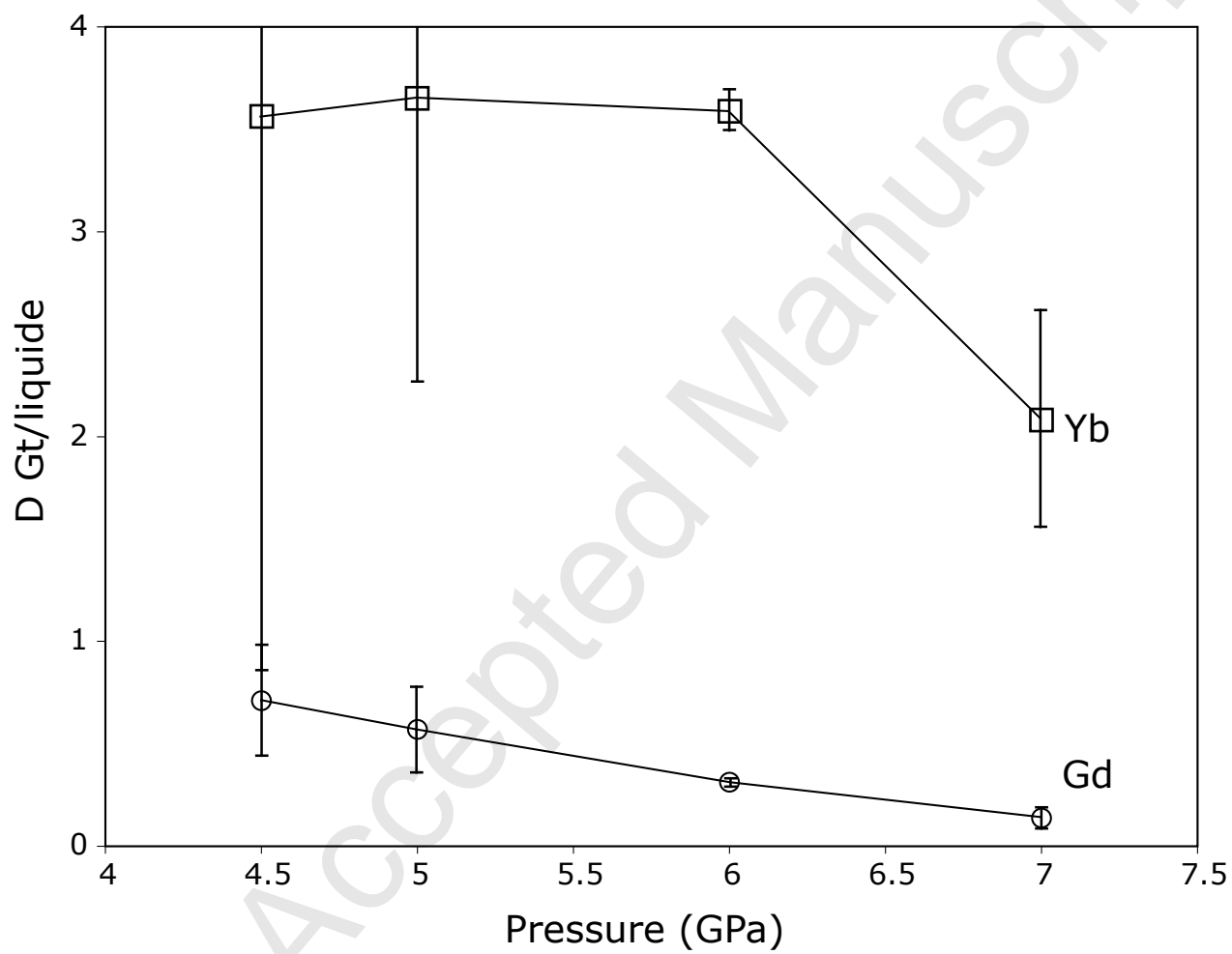


FIGURE 4

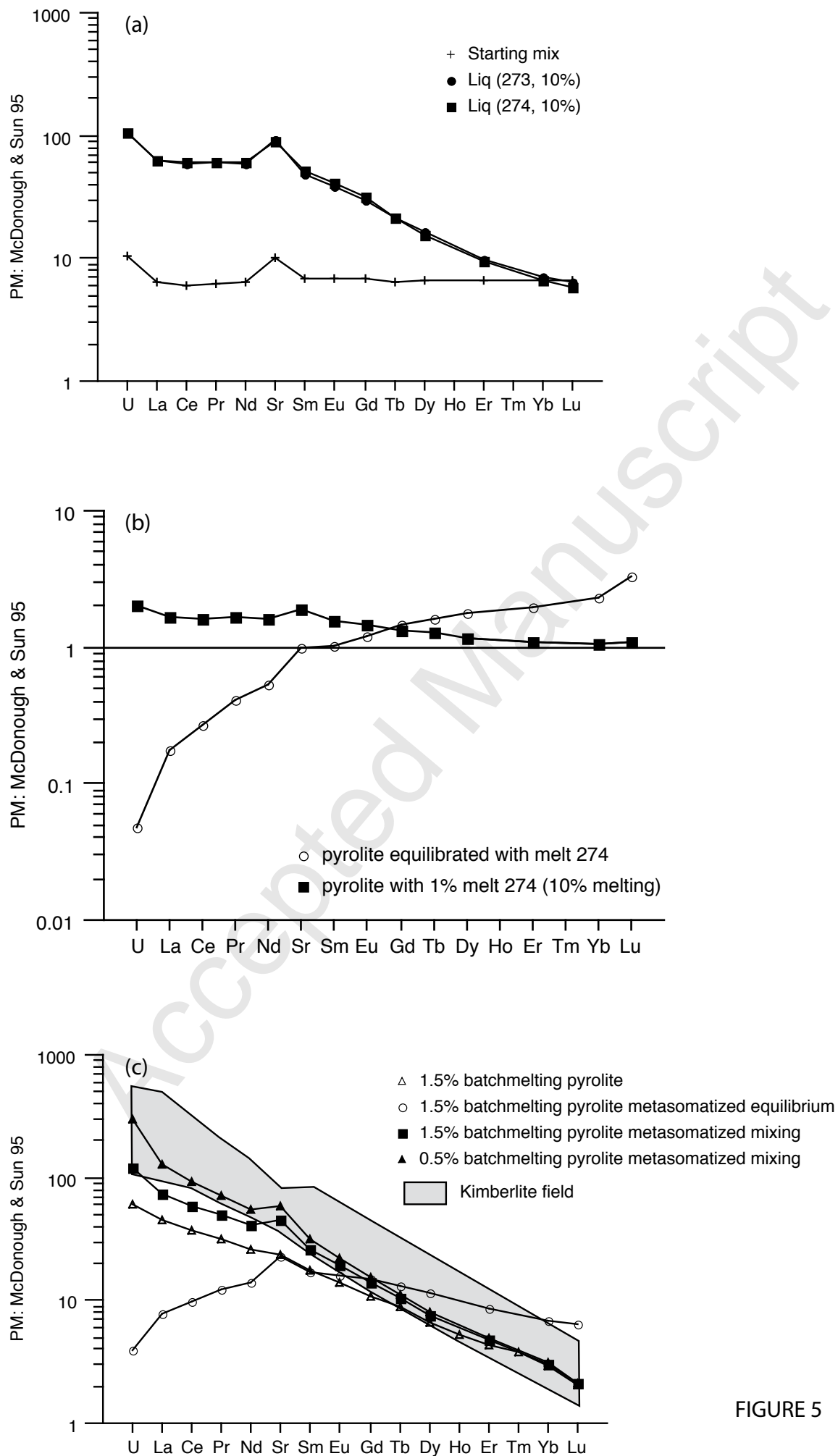


FIGURE 5