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1 **H<sub>2</sub>O-fluid-saturated melting of subducted continental crust facilitates exhumation**  
2 **of ultrahigh-pressure rocks in continental subduction zones**

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10

11 **Abstract**

12 We present two-dimensional numerical models of plate subduction and collision  
13 inspired by the Scandinavian Caledonian orogeny to investigate the possible impact of  
14 continental crust partial melting on the exhumation of Ultra-High Pressure metamorphic  
15 rocks. Three possible reactions were tested: low temperature solidus representing H<sub>2</sub>O-  
16 fluid-saturated partial melting, and two end-member reaction curves for dehydration  
17 melting. Thermo-mechanical effects of partial melting were implemented as (1) a  
18 viscosity decrease as a determined rheologically critical melt percentage was reached  
19 (here 0.1), (2) a change in effective heat capacity and adiabatic heating/cooling  
20 accounting for a latent heat term in the heat equation. Among the 3 tested reactions,  
21 only H<sub>2</sub>O-fluid-saturated partial melting drastically modifies the collision dynamics from  
22 the non-melting reference model holding all other parameters constant. A substantially

23 low general viscosity trunction (here  $10^{17}$  Pa.s) is needed to properly resolve the effect  
24 of partial melting on deep collision processes. Low temperature melting indeed induces  
25 the development of a low viscosity buoyant plume prior to slab detachment, where  
26 migmatites exhume from UHP conditions at rates and with pressure-temperature paths  
27 similar to the natural values acknowledged for the Norwegian Caledonides. High  
28 temperature melting has no drastic influence on early collision dynamics. While positive  
29 buoyancy remains the first order driver for the exhumation of buried continental rocks,  
30 exhumation initiates in these cases with eduction subsequent to slab detachment.  
31 Melting and formation of a migmatite plume can later occur along decompression path  
32 while continental crust undergoes thermal reequilibration at temperatures above 900°C.  
33 Some of the partially molten material can also relaminate in the overriding plate rather  
34 than exhume within the collision zone. Even if minor in terms of amount of magma  
35 produced, H<sub>2</sub>O-fluid-saturated partial melting at UHP conditions could therefore have a  
36 dramatic rheological effect and actually limits continental rocks subduction and  
37 facilitates their exhumation.

### 38 **Keywords**

39 UltraHigh Pressure metamorphism, partial melting, continental subduction

### 40 **Highlights**

- 41 • Partial melting of continental crust influences early subduction collision  
42 dynamics.
- 43 • H<sub>2</sub>O-fluid-saturated partial melting can explain specific exhumation history of the  
44 Norwegian Caledonides migmatite-bearing ultra-high pressure rocks.

45 • Partial melting of the crust and slab detachment both limit continental  
46 subduction.

47

## 48 **1. Introduction**

49 Partial melting is considered as a first order weakening mechanism for continental crust  
50 based on experimental evidences (Rosenberg, 2001). While most studies infer a  
51 rheologically critical melt percentage (RCMP), beyond which migmatites are sensitively  
52 weaker than their protolith, values for this threshold vary from 7 up to 20 percents  
53 according to experiments (Rosenberg and Handy, 2005). In the Scandinavian  
54 Caledonides, some of the (Ultra) High Pressure, (U)HP bearing Western Gneiss (WGR)  
55 show migmatization associated to their exhumation, whereas others do not (Labrousse  
56 et al., 2002; Hacker et al., 2010; Kylander-Clark et al., 2014). Whether partial melting  
57 occurs via dehydration melting (Auzanneau et al., 2006) or at least begun in the  
58 presence of free fluids (Ganzhorn et al., 2014) is a key issue. H<sub>2</sub>O-fluid-saturated melting  
59 indeed could occur as early as the pressure peak assessed for the UHP eclogites  
60 (Labrousse et al., 2011) while dehydration reactions are crossed along retrograde path  
61 only. Sources for aqueous fluids in the 2.5-5 GPa pressure range within a subduction  
62 wedge are actually diverse (nominally anhydrous minerals, mantle hydrous phases  
63 breakdown) and fluid saturated melting is considered as a relevant reaction in these  
64 conditions. First increments of H<sub>2</sub>O-fluid-saturated melting at peak conditions could act  
65 as a first weakening mechanism prone to decouple fertile continental crust from  
66 subjacent dry crust and mantle, hence triggering the exhumation of buoyant partially  
67 molten crust (Labrousse et al., 2011).

68 In this paper, we employ previously designed subduction-collision numerical models  
69 (Duretz et al., 2012a) to simulate the Caledonian collision event and Western Gneiss  
70 UHP province metamorphic history. We evaluate the impact of different partial melting  
71 reactions on the dynamics of UHP rocks exhumation, with emphasis on the exhumation  
72 rates and PT paths of UHP continental units. Possible relations between crustal partial  
73 melting and lithospheric delamination are also discussed.

74

## 75 **2. Geological evidence of relations between UHP metamorphics exhumation and** 76 **partial melting**

77 All UHP metamorphic domains worldwide (Figure 1) share clockwise PT paths with peak  
78 pressure and temperature reached coevally in the coesite stability field. Nevertheless  
79 the shape of these paths and their position relative to relevant melting curves differ. In  
80 the Dabie Shan, Late Triassic- Early Jurassic partial melting occurred in the Central  
81 Dabie Shan migmatite UHP eclogite-bearing dome (Faure et al., 2003) which has a PT  
82 path (CDD, Figure 1) significantly hotter than other UHP units in the area, with an initial  
83 isothermal decompression path portion. In the Kokchetav Massif, Kazakhstan, UHP  
84 gneisses show compositions consistent with melt extraction (Shatsky et al., 1999).  
85 Dehydration melting reaction producing garnet in the coesite stability field is expected  
86 to occur according to documented PT path (Hermann et al., 2001). Zircon domains  
87 attributed to peak and early decompression stages (domain 2, Hermann et al., 2001)  
88 show REE patterns in equilibrium with restitic garnet, which is compatible with partial  
89 melting from the very first stages of exhumation, i.e. peak pressure. In the Erzgebirge,  
90 Bohemian Massif, coesite and diamond-bearing gneisses (Massonne, 2003) were  
91 exhumed from pressures higher than 4.5 GPa. Diamonds included in UHP garnets are

92 interpreted as crystallized from a silicate melt (Massonne, 2003) and polyphase  
93 inclusions are interpreted as hydrous melts trapped in UHP conditions (Ferrero et al.,  
94 2015). In the WGR (Norwegian Caledonides, leucosome abundance is rising from south  
95 (SWGR, Figure 1) where lower pressure UHP eclogites are mostly exhumed within  
96 unmolten gneiss, to north (NWGR, Figure 1) where diamond bearing UHP eclogites are  
97 exhumed within pervasively molten gneiss (Hacker et al., 2010; Kylander-Clark and  
98 Hacker, 2014). Zircon ages from leucosomes overlap with age interval for UHP peak  
99 metamorphism (Gordon et al., 2013). Polycrystalline inclusions in eclogitic rims of  
100 garnets (Ganzhorn et al., 2014) evidence presence of silica-rich fluid at peak conditions  
101 and composition of early leucosomes are compatible with partial melting at UHP  
102 conditions (Labrousse et al., 2011). In the North Eastern Greenland eclogite province  
103 (NEGEP), partial melting of gneiss occurred in the coesite stability field (Lang and Gilotti,  
104 2007) via phengite destabilization. Peak PT estimates are close to phengite dehydration  
105 reaction (Hermann et al., 2006). Partial melting on the retrograde path is also recorded  
106 (Gilotti and McClelland, 2007), and could be due to early retrograde garnet-producing  
107 reactions such as the one experimentally identified by Auzanneau et al. (2006). In the  
108 Sulu Orogen coesite-bearing zircons in gneiss showing evidence of equilibration with a  
109 melt yield ages clustering between 220 and 230 Ma, while peak metamorphism age is  
110 considered as 245-225 Ma in age (Hacker et al., 1998; Liu et al., 2012), meaning that  
111 partial melting occurred along the very first increments of retrograde path and  
112 promoted the exhumation of UHP metamorphics.

113 On the other hand, some UHP terranes did not undergo partial melting. In the Alps, Lago  
114 di Cignana unit, Zermatt-Saas Zone, which is the sole UHP unit with oceanic affinities  
115 (Reinecke, 1998) and Dora Maira Brossasco-Val Gilba unit, which is a kilometre-size  
116 metagranitic unit (Chopin and Schertl, 1999) both exhibit a cold peak metamorphism,

117 and no evidence of partial melting. Southern Dabie Chan complex, which is considered  
118 as an intruded metasedimentary complex unit thrust on top of a UHP free gneiss unit  
119 (Hacker et al., 2000) show interaction with fluids at UHP (Xia et al., 2008) but no direct  
120 evidence of partial melting.

121 Compilation of PT paths from UHP metamorphic units shows that those exhibiting  
122 migmatites cross one experimental solidus curve at or close to their peak pressure, and  
123 that their first exhumation stages systematically occur via isothermal decompression  
124 (Figure 1). This specificity has been explained by adiabatic behaviour during fast  
125 exhumation of the top of the continental panel (Hermann et al., 2001) when lithospheric  
126 slab detaches. No systematic difference in exhumation rate is nevertheless reported  
127 between UHP units with isothermal decompression and early thermal reequilibration.  
128 All combinations between hot/warm/cold and fast or slow exhumations have been  
129 indeed documented (Kylander-Clark et al., 2012; McClelland and Lapen, 2013).

130 In order to assess the possible effects of partial melting reactions in the  
131 subduction/exhumation dynamics of continental crust during early collision, thermo-  
132 mechanical modelling has been performed using different parameterisations for  
133 melting. The Norwegian Caledonides have been chosen as a case study for model set-up  
134 and comparison of synthetic outputs with natural data.

135

### 136 **3. Numerical modelling**

#### 137 **3.1 Set up and partial melting implementation**

138 The model setup, borrowed from Duretz et al. (2012), consists of two continents  
139 separated by a 500 km wide ocean. Simulations were run with the

140 thermomechanical code I2VIS (Gerya and Yuen, 2003, and appendix). Each  
141 lithology was characterized by a temperature-stress dependent visco-plastic  
142 rheology (rheological and flow parameters as listed in table 1). The size of the  
143 model was 4000 x 1400 km and variable grid spacing (1361 x 351 nodes) was  
144 employed to attain 1 km grid spacing in the collision area (Figure A1). We initially  
145 prescribed plate convergence rate at 5 cm/y until 500 km of convergence was  
146 achieved, then let the models be driven by buoyancy forces (e.g. slab pull). All  
147 mechanical boundary conditions were free slip. Surface and bottom temperature  
148 were kept constant, while model sides were subjected to a zero flux condition. In  
149 the simulations presented here, the initial continental crust thickness is 35 km  
150 with a linear geotherm. The oceanic domain has a 7 km thick crust with an initial  
151 thermal structure defined by a half-space cooling solution (40 Myr thermal age,  
152  $10^{-6} \text{ m}^2 \cdot \text{s}^{-1}$  diffusivity). The lithosphere asthenosphere boundary is determined by  
153 the 1573 K isotherm. Upper and lower crustal densities evolve according to the  
154 equation of state; whereas oceanic crust and mantle densities are pre-computed  
155 from phase equilibria assuming basaltic and pyrolitic compositions, respectively.  
156 The upper and lower crust are assumed to be strong (Adirondacks felsic  
157 granulite; Wilks and Carter, 1990) in order to mimic the strong Baltican shield  
158 and maximize the effect of exhumation (Duretz et al, 2012a; Duretz et al, 2012b).

159 Surface processes (erosion and sedimentation) considered as important in  
160 subduction (Pysklywec, 2006) and continental collision (Gerya et al., 2008) have  
161 been fixed to an average value of 1 mm/yr erosion/sedimentation rate. As  
162 demonstrated in Gerya et al. (2008), erosion mainly has an impact on the late



163 evolution of the collision wedge. In this study, we focus on the early stages of  
164 collision and since no quantitative value is known for the erosion rates during the  
165 Scandian collision event, a reasonable constant value has been assigned.

166 Partial melting is parameterized for each lithology (table 2) and three different  
167 continental crust melting reactions were tested (Figure 1): H<sub>2</sub>O-fluid-saturated melting,  
168 as experimentally constrained up to UHP domain (Huang and Wyllie, 1981), low-  
169 temperature undersaturated dehydration melting as determined in the plagioclase field  
170 by Castro et al. (2000) and in the jadeite field by Auzanneau et al., (2006) and eventually  
171 phengite-dehydration melting as determined by Hermann (2002), which constitutes a  
172 high T end-member of the phengite-dehydration melting curves. Estimates of second  
173 critical endpoint (Schmidt et al., 2004; Hermann and Spandler, 2007), above which the  
174 wet solidus is not defined and fluids continuously evolve from aqueous to supercritical  
175 and eventually melts, lie at 5 GPa, meaning that all reactions can be considered at least  
176 up to this pressure. Solidus and liquidus for other lithologies were implemented as  $T =$   
177  $f(P)$  (table 2).

178 In the presented model, partial melting has two important consequences. (1) In the heat  
179 equation, heat capacity is modified according to melt fraction and latent heat of  
180 melting/crystallization is taken into account. (2) Rheology is sensitive to melt fraction.  
181 At any P,T point above the solidus curve, melt fraction F is computed as  $F = [T - T_{sol}(P)] /$   
182  $[T_{liq}(P) - T_{sol}(P)]$ , with  $T_{sol}$  and  $T_{liq}$  the solidus and liquidus temperatures respectively.  
183 For F below (resp. above) RCMP, effective viscosity is the Voigt average (resp. Reuss) of  
184 the protolith viscosity in the same conditions and the general viscosity truncation value.  
185 This implementation is simple enough to be computed for any of the melting scenarios  
186 explored, and is actually a fair approximation of the qualitative strength curves deduced

187 from experimental data (Rosenberg and Handy, 2005). Continental crust crossing its  
188 wet solidus at 3 GPa, 660 °C, reaches RCMP=0.1 at 716°C for the same pressure, and  
189 undergoes a viscosity drop from  $10^{20.3}$  to  $10^{17.9}$  Pa.s (Figure 2M).

190 In I2VIS a minimum viscosity truncation has to be set to limit the adaptative reduction of  
191 time-steps, this minimum viscosity being also attributed to the sticky air layer employed  
192 to mimic the effect of a free surface (Schmeling et al., 2008; Crameri et al., 2012). Since  
193 partial melting implementation induces high viscosity contrasts in our models, a  
194 sensitivity test has been performed to eventually use a  $10^{17}$  Pa.s lower viscosity  
195 threshold in simulations performed to fully observe the impact of different partial  
196 melting reactions (Figure 2 A to L).

197 No density change has been associated to melting here since (1) different melting  
198 reactions may have opposite bulk effects on density (Clemens and Droop, 1998) and (2)  
199 the increase of subducted rocks buoyancy by partial melting enhances development of  
200 buoyant structures (e.g., Gerya et al., 2008) and thus overlaps with rheological effects,  
201 which we want to investigate in this study.

202 We did not take into account the extraction of melt to the surface (i.e. volcanism) since  
203 migmatites in the WGR never exceeded metatexite stage, and therefore unlikely  
204 produced large amounts of melt. No Caledonian peraluminous granite has been  
205 described so far in the WGR or in the allochthonous units overlying them. No  
206 consequent mass transfer therefore occurred due to partial melting, and migmatites can  
207 be modelled as a phase on their own without melt extraction.

208

## 209 **3.2 Results**

210 The sensitivity tests performed to choose the lower viscosity truncation value sufficient to  
211 resolve viscosity contrasts due to partial melting show drastic differences in geometries  
212 and computation times (Figure 2). We ran our model with the H<sub>2</sub>O-saturated melting  
213 reaction and with 10<sup>19</sup>, 10<sup>18</sup> and 10<sup>17</sup> Pa.s viscosity truncations. In the 10<sup>19</sup> Pa.s  
214 truncation case (Figure 2A to D) UHP metamorphic rocks are exhumed as a buoyant  
215 plume within the subduction channel directly to the surface, then forming a UHP nappe  
216 on top of the subducted plate (Figure 2D). In both other cases, exhumation at the plate  
217 interface is stopped at sub-crustal depth (30-40 km) and UHP rocks form a deep core  
218 (Figure 2H and L) that will be further exhumed by late orogenic processes. Viscosities in  
219 the 3 simulations allow to explain this first order difference. In the highest truncation  
220 viscosity case, the subduction interface constitutes an artifactual continuous low  
221 viscosity channel connecting UHP metamorphic rocks directly to the air. They exhume  
222 along this channel due to extreme pressure gradient. In the two other cases, low  
223 viscosity regions in the upper subduction wedge and in the UHP domain are not  
224 connected, and UHP rocks only exhume in the deeper part of the subduction interface.  
225 Considering that the differences between 10<sup>18</sup> and 10<sup>17</sup> Pa.s viscosity truncation cases  
226 were minor, we considered this latest value as sufficient for the comparison of the  
227 impact of different melting reactions.

228 Reference experiment with no crustal partial melting implemented (Figure 3A) develops  
229 a continental subduction beneath a partially hydrated mantle wedge. The geometry of  
230 the hydrated interface is continuous, even though poorly resolved in our models (rising  
231 hydrated "fingers" are only 1 mesh cell large). Continental crust reaches maximum  
232 depth and lithostatic pressure corresponding to the UHP field (depths beyond 100 km)  
233 about 8 Myrs after the lower plate continental margin is involved in the subduction, i.e.  
234 at the same time the lower plate lithosphere starts to detach by necking at the continent-

235 ocean transition (Duretz et al., 2012b). Continental crust is then relaminated in the  
236 overriding plate or exhumed along the plate interface due to lower plate rollback. The  
237 initial stages of this reference model compare both in geometry and timing with the  
238 initial stages of the subduction-eduction model proposed by Duretz and co-workers  
239 (2012a) for the Norwegian Caledonides.

240 Experiments with dehydration melting implemented at low or high temperature (Figure  
241 3C and D) show very similar pattern, with slab detachment occurring about 1 Myr  
242 earlier than in the reference model. Partially molten continental crust then decouples  
243 from the lithospheric mantle, to be relaminated within the upper plate lithosphere. In  
244 the LT dehydration case exhumation of migmatites is eventually achieved at 28 Myr (ie  
245 12 Myr after collision). In the HT dehydration case migmatites are not exhumed and  
246 remain within the lower plate.

247 Experiments with low temperature H<sub>2</sub>O-fluid-saturated partial melting parametrization  
248 show a substantially different pattern (figure 3B and Figure 4). Partially molten  
249 continental crust is exhumed within the subduction interface as soon as 7 Myrs after  
250 collision initiates and acts as a low viscosity decoupling interface between the upper and  
251 lower plates (Figure 3E). They form a pointing up wedge developing into a 20 km wide  
252 crustal plume between the two converging plates. Rocks, with melt percentages up to 24  
253 % have a high turbulence dynamics within this partially molten channel so that markers  
254 initially located away from each other in the downgoing slab eventually end out close in  
255 the partially molten domain (Figure 4C). Widening of the channel and recumbent folding  
256 of the overriding plate maintains the migmatites in lower position until they reach  
257 crustal depth (Figure 4C). Slab detachment is delayed and occurs 9 Myrs after collision.  
258 Early melting and exhumation of the subducted margin explain this difference in slab

259 detachment timing. Early exhumation hence reduces the bulk positive buoyancy of the  
260 continental slab therefore delaying slab detachment.

261 Duration of residence in the eclogite field (Figure 5) has been measured for selected  
262 markers within the distal (A, C, E and G in Figure 5) and proximal continental margin  
263 (B,D,F and H in Figure 5) and range between 13.5 and 14.4 Myrs for reference non-  
264 molten model and dehydration melting models. Only the H<sub>2</sub>O-fluid-saturated model  
265 markers yield shorter residence times for 9.7 and 10.5 Myrs for the proximal and distal  
266 marker respectively. Exhumation rates were computed in all models for the same  
267 markers in the downgoing plate. Involvement of the margin in the subduction is marked  
268 by a 3 Myr negative peak indicating burial with maximum rate between 105 and 142  
269 mm/yr for proximal margin marker and 137 and 173 mm/yr for distal margin marker.  
270 These maximum values are actually computed over very short time intervals (80 ky).  
271 Averages for burial rates calculated over the 15-19 Ma interval yield values between 47  
272 and 79 mm/yr, which should be compared to spontaneous convergence rate in the  
273 experiment ie 28 mm/yr average during subduction. No unequivocal impact of partial  
274 melting on burial dynamics can be inferred from these experiments. While reference  
275 model exhibits an exhumation pulse synchronous with slab detachment at 24 Myr with  
276 maximum exhumation rate about 35 mm/yr, and average exhumation rate over the 22-  
277 32 Myr interval between 6 and 13 mm/yr, models with partial melting implemented  
278 show much more contrasted exhumation histories. LT Dehydration melting scenario  
279 (Figure 5 E and F) shows exhumation of the subducted continental margin triggered  
280 during slab detachment before partial melting occurs, with maximum exhumation rates  
281 up to 14 mm/yr. Reaching of solidus and subsequent RCMP are not correlated with an  
282 increase in exhumation rate; sensible exhumation (maximum value 49 mm/yr) is  
283 achieved later due to slab retreat. In the HT dehydration melting scenario (Figure 5 G

284 and H), exhumation of the subducted continental margin is also triggered during slab  
285 detachment and reaches moderate exhumation rates. No exhumation is associated to  
286 melting and reaching of RCMP. In both cases average exhumation rate over the 22-32  
287 Myr interval range between 4 and 9 mm/yr, i.e. lower values than for the reference  
288 model without melting implemented. In the H<sub>2</sub>O-fluid-saturated melting case,  
289 exhumation rates are much higher than in the other cases. Maximum values compare  
290 with maximum burial rates (145 and 190 mm/yr for proximal and distal margin  
291 respectively), with average values over the 22-32 Myr period between 12 and 18  
292 mm/yr. Most of exhumation is achieved over a short time interval (about 1 to 2.5 Myrs)  
293 prior to slab detachment and immediately after RCMP is reached. Exhumation is  
294 therefore synchronous to subduction, migmatites rising upward in a counterflow  
295 exhumation plume (Figure 2B) while the lower plate is still subducting.

296 Synthetic PT paths, directly comparable with natural data as compiled in Figure 1 were  
297 plotted for the same markers from the distal and proximal subducted passive margin  
298 (Figure 6). Markers initially equally spaced share the same prograde path in the  
299 reference model and the 3 different partial melting explored cases. PT paths for the  
300 internal margin (paths B, F and H, Figure 6) are sensibly identical for the non-melting  
301 reference case and the dehydration melting scenarios with maximum temperature  
302 about 700°C reached during decompression. Paths for external margin also compare in  
303 the reference model and the HT dehydration case, where ultra-high temperatures (UHT,  
304 > 900°C) and a high partial melting rate close to 30 % are reached during relamination  
305 in the upper lithospheric plate (circle 1 on Figure 6). PT paths for H<sub>2</sub>O-fluid-saturated  
306 melting (C and D paths, Figure 6) coevally reach maximum pressure and temperature  
307 (not exceeding 860 and 810°C for the distal and proximal markers respectively) and  
308 then undergo exhumation via isothermal or slightly cooling trajectories. RCMP (ie 0.1

309 melt fraction rate is reached close to peak and maximum melt fractions reached are  
310 about 20 %. Chaotic portions of the path (as along path D at 2.6 GPa, Figure 6) betray  
311 mixing within the partially molten plume. PT path for the external margin in the LT  
312 dehydration case shows an intermediate shape (path E, Figure 6), with maximum P  
313 reached in the non-molten domain, and a change in PT path slope when RCMP value is  
314 reached, with a nearly isothermal decompression at temperature between 910 and  
315 850°C along the 20% melt fraction curve. Apart from external margin markers in the  
316 reference and HT dehydration models (A and G paths), all markers end out noteworthy  
317 in a common PT domain around 2 GPa and 700°C (circle 2 on Figure 6).

318

#### 319 **4. Discussion**

320 Our models show that low temperature (in the 600-700°C temperature range) melting,  
321 i.e. in the presence of free fluid, drastically modifies collision wedge dynamics with the  
322 development of a low viscosity buoyancy-driven exhumation plume early in the collision  
323 history. Exhumation of UHP partially molten continental rocks is facilitated by the  
324 rheological weakening due to partial melting and PT paths of these units share a high  
325 rate isothermal exhumation stage just after peak pressure and temperature are coevally  
326 reached. In order to realistically resolve this process in our numerical models, a low  
327 viscosity truncation is needed, not higher than  $10^{18}$  Pa.s. The exhumation dynamics  
328 produced is considered as more realistic than the one stage nappe-like exhumation  
329 obtained for high viscosity truncation (Figure 2 A to D) for 1- it reproduces the observed  
330 tectonostratigraphy in the WGR: UHP rocks form domes below HP domains rather than  
331 nappes on top of lower grade rocks, and 2- it decouples deep and shallow orogenic

332 exhumation dynamics and explains the two stage PT paths documented in the WGR and  
333 in most of the UHP domains worldwide.

334 Validation of the presented models can be achieved through the comparison of melt  
335 fractions in the migmatites, residence time intervals in the eclogites field, synthetic  
336 burial and exhumation rates and eventually PT paths with natural data from the  
337 Norwegian Caledonides used as a case study.

338 In the WGR, melt fractions observed in the migmatites are not higher than 30%,  
339 resulting in metatexite textures only. Even though these textures do not evidence the  
340 absence of melt mobility, they actually imply that this mobility remained local, and that  
341 no substantial melt extraction occurred. The absence of Caledonian leucogranites can  
342 confirm this assertion. Models computed therefore must also comply with this low melt  
343 mobility hypothesis. Melt fractions achieved by markers within the migmatitic UHP unit  
344 (Figure 4 and 6) reach maximum values close to 30%. They reach 24 % for the  
345 migmatite plume in the H<sub>2</sub>O-fluid-saturated melting model, and 32% in the HT  
346 dehydration partial melting model, where migmatites are relaminated in the upper  
347 plate. The condition of low melt fraction is therefore respected, and the consideration of  
348 migmatites as a rheological phase on their own is valid here. Extraction could actually be  
349 relevantly considered in the high melt fraction relaminated material in the HT  
350 dehydration model. In the case of H<sub>2</sub>O-fluid-saturated melting, the 20 % melt fraction  
351 reached by the migmatites at 850°C and 3 GPa along their retrograde path (Figure 6)  
352 imply 2 wt % H<sub>2</sub>O initially present according to recent experimental data compilations  
353 (Mann and Schmidt, 2015). Even if no definitive numbers are assigned to fluid fluxes in  
354 the deep part of orogens, dehydration reactions in the mantle are prone to produce  
355 fluids in the same proportions. These values are therefore not out of range.



356 Residence times in the garnet-crystallization field have been estimated in the WGR  
357 through paired Lu-Hf and Sm-Nd dating on eclogites (Kylander-Clark et al., 2009 and  
358 references therein). These values can be considered as minimum equivalents of the  
359 residence time in the eclogite facies as computed for our markers. The four meaningful  
360 age pairs in the 390-420 Ma range documented in the WGR give minimum residence  
361 durations of 17 to 9 Myrs for eclogites sampled out of the UHP domains, and 6 Myrs for  
362 the UHP eclogite analyzed in the Sorøyane UHP domain (Kylander-Clark et al., 2009).  
363 Our models give comparable values for residence in the eclogite field (9.7 to 14.4 Myrs).  
364 The coincidence between the shorter residence time computed for the H<sub>2</sub>O-fluid-  
365 saturated melting model and with the shorter Lu-Hf/Sm-Nd age difference measured in  
366 a migmatite-hosted UHP eclogite is to be noted. Nevertheless uncertainties on the age  
367 differences being the same order of magnitude as their values, this fitting may be not  
368 conclusive.

369 Spontaneous burial rate during continental subduction (47 to 79 mm/a) also lay within  
370 the order of magnitude of convergence rate admitted for Laurentia-Baltica convergence  
371 at the onset of the Scandian collision (Torsvik et al., 1996). Instantaneous exhumation  
372 rates derived from the model (Figure 5) have absolute values 1 order of magnitude  
373 higher than ever recorded in high-pressure rocks. The highest values are computed on  
374 very short time steps (down to 32 kyrs) while significant exhumation rates can only be  
375 confidently assessed from natural data over time intervals larger than uncertainties on  
376 radiochronological data, i.e. not less than a few Myrs. Average values from numerical  
377 models yield more realistic estimates. The 4 to 18 mm/yr exhumation rates computed  
378 for external to internal margin markers in the different models compare in order of  
379 magnitude with the range of values deduced from natural data in the NWGR (100 km  
380 exhumation over 5 Myrs, ie 20 mm/yr, Hacker et al., 2007). The models also show that

381 long residence time in the eclogites field do not imply slow subduction or slow  
382 exhumation as argued by some authors in the WGR on the basis of scatter of  
383 crystallization ages for eclogites (Kylander-Clark et al., 2008; Kylander-Clark et al., 2009).  
384 H<sub>2</sub>O-fluid-saturated melting case yields the fastest exhumations with average rates over  
385 10 mm/yr (12 to 18 mm/yr for external to internal margin) for the very first stages of  
386 exhumation. In this prospect, it is worth comparing synthetic data with exhumation  
387 timing and rates in SWGR, where UHP eclogites are exhumed without significant  
388 melting, and NWGR where UHP eclogites are closely associated to migmatites. SWGR  
389 have equilibrated at UHP at 410-400 (Kylander-Clark and Hacker, 2014), while NWGR  
390 UHP eclogites reached peak at 415-420 and reequilibrated at HP at 400 Ma (Terry et  
391 al., 2000). Calculation of exhumation rates (Labrousse et al., 2004) for the different UHP  
392 domains yield slightly lower values (2.7 to 2.3 mm/a) for SWGR than for NWGR (8.1  
393 mm/a) at least for the initial exhumation stages. Although these 10 to 5 Myrs differences  
394 in timing are within radiochronology uncertainties, it seems therefore that SWGR  
395 exhumed later and slower than NWGR. Our model could actually fit these data  
396 considering the proximal marker in our non-molten simulation as representative for  
397 SWGR UHP domain and the distal marker of our H<sub>2</sub>O-fluid-saturated melting model as  
398 representative for NWGR UHP domain. This would imply that NWGR initially had a more  
399 distal position than SWGR, which is possible but not corroborated by any geological  
400 data.

401 Temperature ranges reached by UHP rocks in the models are in good agreement with  
402 natural values and actually constitute a better discrimination criterion between models.  
403 Only H<sub>2</sub>O-fluid-saturated melting scenario produces UHP rocks with peak conditions  
404 and subsequent decompression in the documented 750-850°C range (as in Hacker al.,  
405 2007 and reference therein). Proximal marker exhumed in the non-molten model

406 (Figure 5B) reaches 740°C during retrogression which is slightly higher than what was  
407 recorded in the non-molten SWGR (Labrousse et al., 2004; Kylander-Clark and Hacker;  
408 2014). A marker intentionally taken further inside the margin could actually fit this  
409 value, so that this inadequation is not critical. Other melting scenarios produce higher  
410 temperature trajectories with UHT reached during decompression, while no index of  
411 such extreme crystallization conditions has been recorded so far in the WGR. Peak  
412 pressures reached by the chosen markers in the models (between 3.7 and 5.2 GPa,  
413 Figure 5) fall between the usually documented pressures in the NWGR (maximum 3.8  
414 GPa in Hacker, 2007) from thermo-barometry on phengite-bearing eclogites mainly and  
415 higher values (5.5 to 6.5 GPa) locally computed from garnet peridotites and websterites  
416 included within the NWGR (Vrijmoed et al., 2006, van Roermund et al., 2009). The width  
417 of the exhumation plume developing in the H<sub>2</sub>O-fluid-saturated melting case (~ 20 km  
418 on Figure 3B) is slightly larger than the minimum width of 10-15 km alleged for the  
419 average WGR UHP "unit" based on heat diffusion considerations (Hacker et al., 2007).  
420 While the latest stages of exhumation of the WGR UHP rocks are associated to crustal  
421 dynamics beyond the scope of the present study, their initial exhumation could actually  
422 be due to decoupling and upward rising of a partially molten crustal unit as imaged in  
423 the H<sub>2</sub>O-fluid-saturated model computed here. The structural position of the WGR  
424 migmatite-bearing UHP domains depicted as diffuse boundary domes cropping out  
425 beneath the lesser metamorphosed Baltican crust (Labrousse et al., 2002; 2004; Hacker  
426 et al., 2007) is also reproduced by our model (Figure 4C). The lower scale aspect of the  
427 WGR migmatite-bearing UHP province, i.e. meter to hectometer mafic and ultramafic  
428 bodies included in a highly strained migmatite matrix could actually be inherited from  
429 high rate mixing within such a migmatitic plume (Figure 4B). The km-size areas  
430 described in the WGR as subducted and exhumed without substantial deformation

431 (Hacker et al., 2010) are not reproduced in our model, where the whole migmatite  
432 plume is affected by intense mixing. Differences in protoliths permeability and initial  
433 chemistry argued for explaining their differential reactivities and strength in UHP  
434 conditions are not implemented in our model, and small scale strain pattern within the  
435 migmatitic exhumed unit is probably not relevant. The continental crust modeled here is  
436 indeed homogeneous and reacts instantly and completely to pressure-temperature  
437 changes.

438 The intersection of synthetic PT paths at 2GPa, 700°C (circle 2 on Figure 6) is similar to a  
439 common retrogression domain noticed in the natural data from the different UHP rocks  
440 in the WGR but at slightly lower conditions (1GPa, 600°C, Labrousse et al., 2004). Both  
441 actually have the same geodynamic significance. These PT domains represent the  
442 transition between exhumation in the mantle (at higher pressures) and within the crust  
443 (at lower pressure). In both cases they coincide with a decrease in exhumation rate,  
444 possibly explaining why these conditions are actually recorded by recrystallization in  
445 the metamorphic rocks. The comparison of our models outputs with independent data  
446 from the natural case, which we chose to define their set-up, show that these data are  
447 best reproduced by the H<sub>2</sub>O-fluid-saturated melting model for the NWGR. The  
448 differences between NWGR and SWGR exhumation rates and timing are also  
449 reproduced, considering that proximal markers in the no-melting model are  
450 representative of SWGR.

451 Our H<sub>2</sub>O-fluid-saturated partial melting model is also consistent with previous models  
452 dedicated to exhumation of UHP continental rocks after continental subduction (Warren  
453 et al., 2008; Gerya et al., 2008, Sizova et al., 2012). Our set-up would range in the high  
454 initial viscosity for the continental margin and high weakening factor, meant to model

455 partial melting reactions among others. Our model yields exhumation of rocks from  
456 comparable PT conditions and with a comparable geometry. Our partially molten plume  
457 is an equivalent of the weak plume exhuming in type III models in Warren et al. (2008).  
458 PT paths however differ; models encompassing weakening only due to strain (Warren et  
459 al, 2008) and no partial melting parameterisation fail at reproducing initial isothermal  
460 decompression in PT paths.

461 Relamination of partially molten rocks in the upper plate as exhibited by the HT  
462 dehydration model (Figure 2D, path G on Figure 6) even if not relevant for the  
463 Norwegian Caledonides case has already been suggested as a mechanism for the  
464 development of felsic continental lower crust (Hacker, 2011). Difference between our  
465 models and conceptual model for relamination is the depth of intercalation of lower  
466 plate material (subcrustal in Hacker et al., 2011 vs intra-lithosphere in our models). Rise  
467 of crustal material in our models is not only controlled by the density contrast but also  
468 by the strength of the surrounding mantle (i.e. Stokes' law). The high melt fraction  
469 achieved by this material would actually require the computation of melt extraction to  
470 better resolve this process.

471 A possible link between partial melting of the continental crust in a convergent setting  
472 and lithospheric delamination or slab detachment has already been proposed. Thermal  
473 relaxation of the collision zone and advection of asthenospheric mantle are mentioned  
474 as possible causes for anatexis. Numerical models (Faccenda et al., 2008) show that high  
475 radiogenic heat continental crust promotes slab detachment by heating and subsequent  
476 weakening of the collisional wedge. No study so far pointed at the possible role of  
477 crustal melting on slab detachment. The present study shows that whatever the partial  
478 melting parameterisation implemented if any, slab detachment occurs merely at the

479 same time (i.e.  $\pm 1$  Myr in the collision history). Partial melting of the fertile upper plate,  
480 concerning the upper 10 to 20 km of the lithospheric pile, has no noteworthy impact on  
481 slab detachment dynamics. Even if not at the same scale, slab detachment and partial  
482 melting rather appear as two independent mechanisms prone to limit continental  
483 subduction and trigger the exhumation of UHP continental rocks.

484 The compilation of PT paths for UHP continental units or domains worldwide  
485 systematically show peak conditions close to one of the acknowledged reaction curve for  
486 anatexis and an initial exhumation stage with slightly isothermal decompression (Figure  
487 1). The models presented here were designed to reproduce the Caledonian orogen,  
488 which appears to be a low temperature end-member in terms of thermal state  
489 (Labrousse et al., 2010), but some of our conclusions may have broader implications.  
490 The shape of PT paths for other UHP migmatite-bearing units worldwide may actually  
491 sign their melting at peak conditions. It is indeed expected and visible in our models that  
492 exhuming UHP units induce an upward curvature of isotherms during decompression,  
493 due to 1) high exhumation rates relative to heat diffusion, and 2) heating of the  
494 surrounding rocks by latent heat dissipation during recrystallization. The reason why  
495 the relevant reaction changes with orogen must reside in the initial chemical  
496 composition of the subducted continental crust.  $H_2O$ , and  $K_2O$  contents at first order,  
497 controlling the amount of mica available for melting (Massonne, 2009). The LT  
498 dehydration model with maximum temperatures about  $900^\circ\text{C}$  reached by the subducted  
499 margin, initial exhumation triggered by slab detachment and fast exhumation lately  
500 facilitated by retrograde partial melting show striking similarities with the conceptual  
501 model proposed for the Kokchetav UHP massif (Hermann et al., 2001). The peak PT  
502 conditions coinciding with HT dehydration curve (Figure 1) could also correspond to  
503 reaching of high RCMP isopleth related to LT dehydration solidus. Proper modelling

504 with set-up inspired from Kokchetav would be necessary to further investigate this  
505 hypothesis. Also our models were not designed to test the impact of several melting  
506 reactions occurring throughout the prograde stages. In reality, it might be possible that  
507 polybaric melting results in a finite product marked by dehydration melting reactions,  
508 although the very first increments of melting were associated to H<sub>2</sub>O-fluid-saturated  
509 partial melting.

510

## 511 **Conclusion**

512 Even though H<sub>2</sub>O-fluid-saturated melting is often considered as a second-order process  
513 in terms of magma production, the present study tends to show a possible rheological  
514 first order impact of this reaction in the subduction-collision dynamics, in the  
515 Norwegian Caledonides case study at least. H<sub>2</sub>O-fluid-saturated partial melting as  
516 parameterized here, and with all other parameters kept equal, actually limits  
517 continental subduction and facilitates upward flow of partially molten rocks early in the  
518 collision history, i.e. prior to any slab detachment. Migmatites develop into an exhuming  
519 plume, which acts as a decoupling layer separating the two converging plates. In this  
520 buoyancy driven plume, UHP migmatites exhume at rates comparable to their burial  
521 rate. Synthetic pressure-temperature-time paths produced are very similar to natural  
522 data from the Caledonides, with high initial exhumation rates and isothermal  
523 decompression following coevally reached peak pressure and temperature. The  
524 proposed triggering effect of H<sub>2</sub>O-fluid-saturated melting on exhumation from field and  
525 experimental perspectives (Labrousse et al., 2011, Ganzhorn et al., 2014) is shown to be  
526 also thermo-mechanically sound. Other possible melting reactions for continental crust,  
527 i.e. dehydration melting, do not drastically change subduction-collision dynamics. Slab

528 detachment remains the first order process limiting continental subduction in these  
529 cases. They nevertheless could be key reactions in explaining the exhumation of warmer  
530 migmatite-bearing UHP units in collisional orogens, which also show high rate initial  
531 isothermal decompression and peak PT conditions close to higher temperature solidus  
532 curves for continental material.

533

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541

### 542 **Appendix : Numerical methods and set-up**

543 The presented numerical simulations were carried out using the code I2VIS (Gerya and  
544 Yuen, 2003). The Stokes equations, the incompressible mass conservation equation, and  
545 the energy conservation equation, take the form of:

$$546 \quad \frac{\partial \tau_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} = \rho g_i$$

$$547 \quad \frac{\partial v_i}{\partial x_i} = 0$$

$$548 \quad \frac{\partial}{\partial x_i} \left( k \frac{\partial T}{\partial x_i} \right) = \rho C_p \frac{DT}{Dt} - H$$



549

550 where  $\tau_{ij}$  is the deviatoric stress tensor,  $P$  is the pressure,  $\rho$  is the density,  $g_i$  is the  
551 gravitational acceleration,  $v_i$  is the velocity vector,  $k$  is the temperature-dependent  
552 thermal conductivity (table 1),  $T$  is the temperature,  $C_p$  is the heat capacity, and  $H$   
553 corresponds to heat sources (radioactive heat production, shear heating, adiabatic  
554 heating, and latent heat). These equations are solved with the finite difference/marker-  
555 in-cell method (see Gerya, 2010; Duretz et al, 2011 for technical aspects). The diffusive  
556 fluxes and production terms are discretized on a 2D Eulerian staggered grid, whereas  
557 the advective fluxes are treated on Lagrangian particles.

558 The rheology is visco-plastic taking into account contributions of diffusion-dislocation  
559 creep, exponential creep (mantle lithosphere only), and Mohr-Coulomb rheology.  
560 Viscosities corresponding to each flow mechanism are computed as follows:

561 
$$\eta_{\text{viscous}} = \eta_0^{\frac{1}{n}} \dot{\epsilon}_{II}^{\frac{1-n}{2n}} e^{\left(\frac{E_A + PV_A}{nRT}\right)}$$

562 
$$\eta_{\text{exponential}} = \left\{ 2A_{\text{Peierls}} e^{\left(\frac{E_A + PV_A}{RT}\right)} e^{\left(1 - \left(\frac{\tau_{II}}{\sigma_{\text{Peierls}}}\right)^k\right)^q} \right\}^{-1}$$

563 
$$\eta_{\text{Mohr-Coulomb}} = \frac{C + P \sin \varphi}{2\dot{\epsilon}_{II}}$$

564

565 The reference viscosities ( $\eta_0$ ), stress exponents ( $n$ ), activation energies ( $E_A$ ), activation  
566 volumes ( $V_A$ ), the cohesion ( $C$ ), and the internal friction angle ( $\varphi$ ) are listed in table 1  
567 and  $R$  is the gas constant. The parameters of exponential creep are taken as  $A_{\text{Peierls}} =$   
568  $10^{7.8} 10^{-12}$  Pa-2s-1 and  $\sigma_{\text{Peierls}} = 9.1$  GPa. The second deviatoric stress ( $\tau_{II}$ ) and strain  
569 rate ( $\dot{\epsilon}_{II}$ ) invariants are expressed as

570 
$$\tau_{II} = \sqrt{\tau_{ij}\tau_{ij}}$$

571 and

$$572 \quad \dot{\epsilon}_{II} = \sqrt{\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}}$$

573 The flow rule that produces to the minimum viscosity is defined as the dominant  
574 deformation mechanism:

$$575 \quad \eta_{\text{eff}} = \min (\eta_{\text{viscous}}, \eta_{\text{exponential}}, \eta_{\text{Mohr-Coulomb}})$$

576 and the stress-strain rate relationship is subsequently expressed as:

$$577 \quad \tau_{ij} = 2 \eta_{\text{eff}} \dot{\epsilon}_{ij},$$

578 where  $\eta_{\text{eff}}$  is the effective viscosity, and  $\dot{\epsilon}_{ij}$  corresponds to the strain rate tensor. The  
579 rheological impact of melting is described in the main text.

580 The density of mafic (gabbroic composition) and ultramafic (pyrolitic composition)  
581 material phases is temperature and pressure dependent. Densities are evaluated at each  
582 timestep from look-up tables that are precomputed via Gibbs energy minimization in the  
583 CaO-FeO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> system (Gerya et al, 2004). The density of felsic materials obey  
584 the following equation of state:

$$585 \quad \rho = \rho_0 [1 - \alpha(T - 298.5)(1 + \beta(10^{-8}P - 10^{-3}))].$$

586 The reference density ( $\rho_0$ ) of the upper and lower continental crusts are 2700 kg/m<sup>3</sup>  
587 and 2800 kg/m<sup>3</sup> respectively. The isothermal compressibility is set to 0.5 10<sup>-3</sup> kbar<sup>-1</sup>  
588 and the thermal conductivity is set to 1.5 10<sup>-5</sup> K<sup>-1</sup>.

589 The free surface is approximated by a sticky air approach (Crameri et al, 2012). The air  
590 layer is 20 km thick and its viscosity is set to the lower viscosity cut-off (see figure 2).

591 Surface processes are modeled using a gross-scale erosion sedimentation law (Gerya,  
592 2010) and the topography ( $h$ ) evolves according to:

593 
$$\frac{\partial h}{\partial t} = v_y^h - v_x^h \frac{\partial h}{\partial x} - \dot{e} + \dot{s},$$

594 where  $v_x^h$  and  $v_y^h$  corresponds to the velocity components of the model's surface,  $\dot{e}$  and  $\dot{s}$   
595 are prescribed erosion and sedimentation rates (0.1 mm/y).

596

597 **Table & Figure captions**

598 **Table 1:** Thermal and mechanical parameters employed in the numerical simulations:  $k$   
599 is the temperature dependent thermal conductivity expressed as a function  $T$ , absolute  
600 temperature in K,  $H_r$  is the radiogenic heat production,  $C_p$  is the specific heat capacity,  
601  $\sin(f)$  is the friction coefficient,  $C$  is the cohesion,  $h_0$  is the reference viscosity,  $n$  is the  
602 stress exponent,  $E_a$  is the activation energy, and  $V_a$  is the activation volume. sediments  
603 and upper oceanic crust (UOC) modeled as wet quartzite (1, Ranalli and Murphy, 1987),  
604 upper and lower continental crust (UCC, LCC) modeled as felsic granulite (2, Wilks and  
605 Carter, 1990), lower oceanic crust (LOC) modeled as anorthosite (3, Ranalli and Murphy,  
606 1987), dry mantle (Dry M.) modeled as dry dunite (4, Chopra and Paterson, 1984),  
607 serpentinized and wet mantle modeled as wet olivine (5, Ranalli, 1995).

608 **Table 2 :** Various solidus employed in the simulations.  $H_2O$ -fluid-saturated, Low  $T$   
609 dehydration and high  $T$  dehydration were used to simulate sediment and upper  
610 continental crust melting. Wet basalt melting was employed for mafic lithologies (lower  
611 continental crust, oceanic crust). Dry and Wet mantle solidus were used for melting of  
612 dry and hydrated ultramafic lithologies.

613 **Figure 1:** PT-paths for UHP metamorphic units and solidus curves used in numerical  
614 models for the continental crust.  $H_2O$ -fluid-saturated solidus from (Huang and Wyllie,  
615 1981), low temperature dehydration-melting solidus from (Auzanneau et al., 2006) and  
616 (Schmidt et al., 2004) and high temperature dehydration melting as experimentally  
617 defined by Hermann et al. (2002). PT-paths for South and North Western Gneiss Region  
618 (SWGR, NWGR Labrousse, 2004), North East Greenland Eclogite Province  
619 (NEGEP, McClelland and Lapen, 2013), South Dabie Chan (SDC) and Central Dabie Dome  
620 (CDD) compiled in Faure et al. (2003), Sulu Orogen (SO, Chen et al., 2013), Kokchetav

621 massif (KM, Hermann et al., 2001), Erzgebirge (ERZ, Massonne, 2003) ),Dora Maira (DM,  
622 Chopin and Schertl, 1999) and Lago di Cignana (LdC, Reinecke, 1998). Coesite and  
623 eclogites stability field as in (Liou et al., 2004).

624 **Figure 2:** Effect of general viscosity truncation value used for computation. A to D :  
625 viscosity truncation  $10^{19}$  Pa.s, E to H : viscosity truncation  $10^{18}$  Pa.s and I to L : viscosity  
626 truncation  $10^{17}$  Pa.s. M: viscosity drop implementation due to continental crust partial  
627 melting in the designed model (curve drawn for H<sub>2</sub>O-fluid-saturated melting at P = 3  
628 GPa, strain rate =  $10^{-15}$  s<sup>-1</sup>). Phase colors legend precised in Figure 3.

629 **Figure 3:** Model results 8 Ma after initiation of collision. A, without any melting effect  
630 implemented, B, with H<sub>2</sub>O-fluid-saturated melting, C with low temperature dehydration  
631 melting and D with high temperature dehydration melting. E: viscosity field for H<sub>2</sub>O-  
632 fluid-saturated melting experiment. u/l..c.c./o. upper/lower continental/oceanic crust,  
633 serp./hydr. l.m. serpentinized/hydrated lithospheric mantle, a. m. asthenospheric  
634 mantle. Numbers for rheologies 1 to 5 refer to table 1, 6 & 7 computed according to table  
635 2 equations.

636 **Figure 4:** Time evolution of H<sub>2</sub>O-fluid-saturated partial melting model. The color code  
637 for phases is similar to figure 3. Isotherms in degree Celsius. Melt fraction is color-coded  
638 for partially molten crust in insets. Only 800°C isotherm (dashed line) is reported there.  
639 Arrows on sketches indicate shear senses and flow directions within the subduction  
640 zone.

641 **Figure 5:** Vertical velocity-time curves for 2 markers within the subducted continental  
642 margin (circles and squares refer respectively to circle and square on inset I). Change  
643 from dashed to continuous line represents solidus crossing, simple black arrows show

644 when RCMP is reached. Grey rectangles show residence time interval in the eclogites  
645 field as delimited in Figure 5. Arrow endings for A and G mean these markers did not  
646 exhume from eclogites field. White double arrows indicate slab detachment timings for  
647 each experiment along the paired advection-time curves. l.m. lithospheric mantle, c.c.  
648 continental crust, o.c. oceanic crust.

649 **Figure 6:** Pressure-temperature paths for markers initially located in the lower plate  
650 margin (inset I Figure 4) and different melting reactions implemented. A to H refer to  
651 curves in Figure 5. Iso-melt fraction curves are drawn for each experiment. Prograde  
652 paths, superposing each other have been cut for readability purposes. Coesite and  
653 eclogites stability field as in Liou et al. (2004).

654 **Figure A1:** Initial model configuration. The different colors correspond to various  
655 material phases (see legend figure 3) and the red square indicates the areas of high  
656 numerical resolution. The initial plate velocity is imposed within the plates (black  
657 arrows). The continental plates subsequently decouple from the lateral box sides during  
658 convergence. The isotherms (white lines) are plotted for each 400° C increment.

659

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- Partial melting of continental crust influences early collision dynamics.
- H<sub>2</sub>O-present partial melting is the most sensible reaction.
- Crustal partial melting and slab detachment both limit continental subduction.
- Melting at peak conditions may have occurred in most of UHP domains.

<b>Material</b>	<b>k(T)</b> (W/m/K)	<b>H<sub>r</sub></b> (W/m <sup>3</sup> )	<b>C<sub>p</sub></b> (J/kg/K)	<b>sin(φ)</b>	<b>C</b> (MPa)	<b>η<sub>0</sub></b> (Pa <sup>n</sup> .s)	<b>n</b>	<b>E<sub>a</sub></b> (J/mol)	<b>V<sub>a</sub></b> (m3)
<b>Sediments</b>	0.64+	1.50e-6	1000	0.15	1	1.97e17	<u>2.3</u>	1.54e5	0.8
	807/(T+77)								
<b>U. O. C.</b>	0.64+	0.25e-6	1000	0.00	20	1.97e17	2.3	1.54e5	0.8
	807/(T+77)								
<b>U. C. C.</b>	0.64+	1.00e-6	1000	0.45	20	4.97e20	3.1	<u>2.43e5</u>	1.2e-5
	807/(T+77)								
<b>L. C. C.</b>	1.18+	0.25e-6	1000	0.45	20	4.97e20	3.1	<u>2.43e5</u>	1.2
	474/(T+77)								
<b>L. O. C.</b>	1.18+	0.25e-6	1000	0.60	20	4.80e22	3.2	2.38e5	<u>0.8</u>
	474/(T+77)								
<b>Dry M.</b>	0.64+	2.20e-8	1000	0.60	40	3.98e16	3.5	5.32e5	0.8
	807/(T+77)								
<b>Serp &amp; Wet M.</b>	0.64+	2.20e-8	1000	0.15	1	5.01e20	<u>4.0</u>	4.70e5	0.8
	807/(T+77)								

table 1

		<b>T (K)</b>
<b>For the upper continental crust and sediments</b>	<b>Water-present melting</b>	866.65+ 28.65/(P+0.097)at P< 2.2 GPa 783+43.47*P at P≥ 2.2GPa
	<b>Low T dehydration melting</b>	916.65+ 8.58/(P+0.035)at P< 0.41 GPa 927.85+13.47*P +17.1*P <sup>2</sup> -2.7*P <sup>3</sup> at P≥ 0.41 GPa
	<b>High T dehydration melting</b>	971+ 7.73/(P+0.0003)at P< 0.46 GPa 947+89.55*P -9*P <sup>2</sup> +6.142*P <sup>3</sup> at P≥ 0.46 GPa
	<b>Liquidus</b>	1262+90*P
<b>For lower continental crust and oceanic crust</b>	<b>Solidus</b>	972.6-70.36/(P+0.354)+77.81/(P+0.354) <sup>2</sup> at P< 1.6 GPa 935.4+3.486*P+6.243*P <sup>2</sup> at P≥ 1.6 GPa
	<b>Liquidus</b>	1423.15+105*P
<b>For ultramafics</b>	<b>Solidus for wet peridotite</b>	1239.8+49.76/(P+0.323) at P< 2.4 GPa 1266.3-11.84*P+3.503*P <sup>2</sup> at P≥ 2.4 GPa
	<b>Solidus for dry peridotite</b>	1393.8+132.9*P-5.104*P <sup>2</sup> at P< 1 GPa 2212.4+30.8*(P-1) at P ≥ 1 GPa
	<b>Liquidus</b>	2017.15+114*P

table 2















