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1 **Rheological implications of extensional detachments : Mediterranean and**  
2 **numerical insights.**

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11

12 **Abstract**

13 The Mediterranean realm shaped by extensive back-arc extension after multiple  
14 collisions between Europe and isolated continental blocks is the second densest  
15 occurrence of Metamorphic Core Complexes (MCC) after the North American Cordillera.  
16 The present review aims at determining the factors controlling detachment system  
17 development during post-orogenic extension in the continental lithosphere. 26 different  
18 detachment zones over 23 different localities are systematically described and show  
19 that MCC structures can be positioned in a three end-member classification. Beside the  
20 high-temperature lower plate end-member, most of the time considered as the  
21 archetypical MCC type, two cold end-members can be identified: a first one preserving  
22 high pressure metasedimentary units in the vicinity of the detachment and represented  
23 by numerous cases in the Mediterranean and a second theoretical one with strain  
24 localized at the bottom of a resistant upper crust. These two end-members show a strain

25 pattern guided by inheritance more than thermal state. The relevance of this three end-  
26 member classification is then tested via numerical modelling of extension of layered  
27 continental lithosphere. The compilation of our results and recently published similar  
28 models show that over a 100 of different simulations the most critical parameter for  
29 developing a MCC is the intra-crustal strength contrast ( $S_{\max}/S_{\min}$ , where  $S_{\max}$  and  $S_{\min}$   
30 stand respectively for the maximum – at the brittle-ductile transition – and minimum –  
31 at the Moho – crustal strength). This contrast has to be risen only by 1 order of  
32 magnitude to change extension mode. Below a critical value of 1000, narrow and wide  
33 rifts develop. MCCs with a perennial detachment are produced for values slightly higher  
34 than 1000, while higher intra-crustal strength contrasts systematically promote double-  
35 domes MCCs, with strain progressively relocalizing within the lower plate. Inherited  
36 layering is as efficient as temperature for raising this intra-crustal strength contrast  
37 while strain rate is actually a second order parameter. Metamorphic reactions are also  
38 to be considered as a first order process in localizing detachments.

39

## 40 **1. Introduction**

41 Even if the seminal comprehensive descriptions of Metamorphic Core Complexes  
42 (MCCs) in the American Cordillera mentioned lower plates constituted of gneiss and  
43 intruded by granites (Snake Range, Miller et al. (1983), Whipple Mountains, Davis et al.  
44 (1986), the actual definition of MCCs: « *Cordilleran metamorphic core complexes appear*  
45 *to be bodies from the middle crust that have been dragged out from beneath fracturing*  
46 *and extending upper crustal rocks, and exposed beneath shallow-dipping (normal slip)*  
47 *faults of large areal extent* » (Lister and Davis, 1989) refers to rocks exhumed from the  
48 middle crust whatever their thermal history. The fundamental property of this middle  
49 crust resides in its ability to flow laterally toward the forming dome (Block and Royden,  
50 2010), to accommodate stretching of the upper plate and preserve a relatively flat Moho  
51 (Figure 1). Even though thermal reequilibration can induce weakening of the lower  
52 crust (Buck, 1991) and increases its ability to flow below a relatively strong upper crust,  
53 a similar strength profile can also be inherited from pre-extension evolution of the  
54 continental crust and promote development of the original structure leading to the  
55 formation of MCCs: the detachment (Davis, pers comm. ). In order to unravel the  
56 rheological meaning of detachments, we propose here a review of extensional shear  
57 zones described as post-orogenic detachments in the Mediterranean realm. These  
58 examples share a common tectonic setting and often affect comparable lithotectonic  
59 successions. A three end-members typology is proposed with high-temperature MCCs  
60 (HT end-member) as one end-member, and two cold MCC end-members with a weak  
61 middle crust due in most cases to the interlayering of high-pressure metasedimentary  
62 nappes during syn-orogenic wedge building (HP end-member). Even if not met in  
63 natural cases, a second theoretical cold end-member can be defined by the presence of a  
64 strong upper crust also due to nappe stacking or even obduction (HS end-member). In

65 both cases, the inherited tectonostratigraphy is responsible for a sensible exaggeration  
66 of the intra-crustal strength contrast.

67 Fully coupled thermo-mechanical modeling experiments allow testing these three end-  
68 member typologies and determining the critical intra-crustal strength contrast for the  
69 perennial development of a detachment zone and building of a dome at the expense of  
70 the lower plate. The comparison of natural cases classification and model-based  
71 categories demonstrates the mechanical feasibility of all natural combinations and  
72 possibly explains how long-lasting multiple detachment systems such as Menderes,  
73 Turkey or North Cycladic Detachment System, Cyclades, Greece, evolve.

74

## 75 **2. Mediterranean Extensional Detachments**

76 Metamorphic Core Complexes are described in the Mediterranean within three large  
77 extended domains (Carpathians and Pannonian basin, Western Mediterranean from  
78 Alboran to Tyrrhenian, and Aegean domain *sensu lato*) subject to slab retreat during  
79 Alpine orogeny (Jolivet et al., 2009a). Most of them are now located in continental  
80 domains with Moho depths between 25 and 20 km, but not systematically in flat Moho  
81 areas (Figure 2). Betic Cordilleras and Menderes are, for instance, on top of dipping  
82 Moho with about 10 km vertical drop across them (Karabulut et al., 2013 for instance for  
83 Menderes). Metamorphic rocks exhumed within Mediterranean core complexes show a  
84 wide scatter in pressure-temperature conditions (Figure 3 and Table 1). Dating peak  
85 metamorphic conditions and detachment mylonites or relative dating of the detachment  
86 activity allow deciphering between rocks mostly exhumed during pre-extensional  
87 thickening stages from rocks exhumed during extensional dynamics leading to the MCC

88 formation. Furthermore, the presence in the footwall of detachments of lower grade  
89 rocks structurally below the High Temperature and Low Pressure (HT-LP) or High  
90 Pressure and Low Temperature (HP-LT) rocks is also a criterion to attribute the first  
91 stages of the exhumation of the latter to syn-orogenic nappe stacking rather than  
92 lithospheric extension. Nappe stacking is indeed the sole process prompt to thrust high-  
93 grade metamorphic units on top of a preserved lower grade unit. For instance, the  
94 equilibration of the Ichendirene HP-granulite bodies (BB1, Figure 3) included in the  
95 lower grade Beni Bousera kinzigites (BB2, Figure 3) in Morrocco relates to stages earlier  
96 than the exhumation of the massif itself (Michard et al., 2006).

97 Even when removing HP-LT peaks inherited from pre-extensional stages and preserved  
98 during dome formation, the scatter of PT conditions sampled by detachments footwalls  
99 remains high. Noteworthy, some core complexes such as the one described in the Catena  
100 Costiera, Calabria, Italy (CC on Figure3, Rossetti et al. (2004), or in the Northern  
101 Cyclades (Tn, Tinos, Jolivet et al. (2010b) and references therein) show « cold »  
102 footwalls, with conditions theoretically uncompatible with the development of  
103 detachment systems, according to numerical models (Tirel et al., 2008). The rheological  
104 meaning of such « cold MCCs » (Huet et al., 2010), is the purpose of our study.

105

## 106 **2.1 Tectonostratigraphic positions**

107 The systematic description of tectonostratigraphic positions of Mediterranean  
108 detachments allows highlighting the preferred decoupling levels within the continental  
109 crust during its extension (Table 1 and 2). Among relevant characteristics, some were  
110 considered necessary for defining a detachment zone in our compilation: the presence of

111 mylonites along the extensional fault system, in order to compare structures initiated  
112 below the ductile-brittle transition only, and a clear extensional setting, in order to  
113 compare post-orogenic extensional detachments only, and avoid exhumation shear zone  
114 (syn-orogenic detachments in Jolivet et al., 2003). Some characteristics differ from a  
115 detachment system to another: the lithology and PT conditions recorded by their  
116 footwall (patterns and color code on Figure 4), the lithology of their hangingwall, as well  
117 as the presence of plutons, with mantle or crustal origin. The presence of migmatites in  
118 the footwall has also been systematically noted. The initiation context of the exhumation  
119 process has also been checked : syn-orogenic for rocks partly exhumed during  
120 convergence, post-orogenic for rocks exhumed only during post-orogenic extension. The  
121 systematic documentation of those characteristics (Figure 4 and Table 2) shows a  
122 distribution of natural cases with three theoretical end-members, between which they  
123 can be positioned in a triangular diagram (Figure 5). The position of each case was  
124 deduced from the relative weighing of high temperature criterion (migmatites or high  
125 temperature lower plate or syn-kinematic intrusives), low temperature criterion  
126 (preservation of HP-LT paragenesis in the footwall) or high strength upper plate  
127 criterion listed in table 2. As will be discussed further, mediterranean natural cases plot  
128 within and along edges of the proposed ternary plot. While the HT and HP end-embers  
129 are represented by numerous natural cases, no case is found on the HS end-member.  
130 None of the documented examples indeed has a strong upper plate as single criterion.

131

### 132 **2.1.1 Detachments on top of High-Temperature lower plates**

133 Some of the Mediterranean detachments cap high-temperature lower plates, with HT-LP metamorphic  
134 rocks showing syn-extensional pervasive amphibolite facies metamorphism, possibly associated to  
135 migmatization, and/or syn-kinematic granite intrusion.

136 The Kabylia Detachment, Greater Kabylia, Algeria (KaD, table 1C), is located between Permian phyllites  
137 in its hanging-wall and amphibolite facies micaschists and orthogneisses in its footwall (Saadallah and  
138 Caby, 1996). Migmatitic textures, syn-tectonic aplitic dikes and late peraluminous granite intrusions  
139 indicate a temperature above the computed solidus (i. e. 700°C) during the exhumation of paragneiss and  
140 micaschists forming most of the lower plate. Rheological contrast between upper and lower plate is  
141 mostly due to contrasted thermal histories. The Edough massif (Ed, table 1D) in Algeria shows a similar  
142 structure (Caby et al., 2001) on top of various high grade metamorphic rocks reaching anatexis and  
143 intruded by syn-kinematic leucogranites. Fragments of sublithospheric mantle were incorporated within  
144 crustal stack and now constitute rheologically constrained inclusions (Bosch et al., 2014). Even if  
145 exhumation ages in the Edough massif are younger than in Greater Kabylia (Bruguier et al., 2009), both  
146 extensional domes actually built from the same crustal stack during Alboran slab retreat or later slab  
147 related dynamics.

148 On Elba island, Italy, even though the cataclastic low angle normal Zuccale fault zone accommodated  
149 substantial extension (Colletini and Holdsworth, 2004), the proper detachment zone (Capanne  
150 Detachment Zone, Table 1H), exhuming the lower Tuscan unit greenschists, is located at the top of the  
151 syntectonic Monte Capanne granodiorite (Daniel and Jolivet, 1995). Mylonites are described in its  
152 thermometamorphic aureole and localization of strain is related to its intrusion.

153 The Lubenik Line (LL, table 1L), in the Central Western Slovakian Carpathians and the Zaouia Fault Zone,  
154 in the Beni Bousera massif, Morocco, even though both considered as structures inherited from orogenic  
155 stages (Janák et al., 2001; Michard et al., 2006) also belong to the High Temperature detachments class. In  
156 the Lubenik Line footwall, the Veporic unit is intruded by the syn-kinematic Rochovce granite and  
157 mylonites localized strain at its top (Janák et al., 2001). In the Beni Bousera massif, amphibolite facies  
158 metamorphism in the Fillali schists and associated kinzigites indeed indicate temperatures higher than  
159 800°C for their peak conditions, and leucogranite intrusions evidence partial melting of the lower plate  
160 (Michard et al., 2006).



161 HT lower plate detachment class is therefore represented by detachment zones that exhume HT  
162 metamorphic rocks, which reached peak P-T conditions just before exhumation or lower plate intruded by  
163 plutons, responsible for local thermal weakening and strain localization. This mechanism is discussed in  
164 section 4.2.3.

165

### 166 **2.1.2 Detachments within interlayered high-pressure metamorphic nappe**

167 Most of the "Cold MCC" (Huet et al., 2010) exhume HP-LT metamorphic rocks in their footwall (Figure 3).  
168 Even though most of these HP-LT rocks were actually partly exhumed prior to the proper detachment  
169 tectonics related to extension, their preservation from thermal reequilibration during ultimate  
170 decompression implies that no significant thermal reequilibration occurred during MCC formation.

171 Detachment faults described in the Betics, Spain (Mecina-Filabres Shear Zone, MSFZ, table 1A), on Syros,  
172 Greece (Vari detachment, VaD, table 1P), in southern Italy (Catena Costiera and Aspromonte, CatCo and  
173 Asp, table 1E & F), in Peloponnese, Greece and Crete (Cretan detachments 1 and 2, CrD1 and CrD2, table  
174 1U & V) and also in the Rechnitz window, Austrian Alps (ReW, table 1K) belong to this category.

175 The Mecina-Filabres Shear Zone develops as mylonites and gouges during extensional reactivation of a  
176 former thrust (Platt et al., 1984). Eclogites are preserved within the Nevado-Filabrides lower plate, and  
177 most of the syn-tectonic metamorphism is actually greenschist facies (Augier et al., 2005b). No drastic  
178 lithological contrast is observed between the upper plate Alpujarrides phyllites, quartzites and graphitic  
179 micaschists and the exhumed Bedar-Macael graphitic schists.

180 The Vari detachment, Syros, exhumes the Cycladic blueschist metasedimentary unit dominated by schists  
181 and marble preserving eclogite to blueschist paragenesis, (Trotet et al., 2001), beneath the Vari and  
182 "Upper unit" metabasites gneiss and serpentinites which experienced earlier HT reequilibration (Jolivet  
183 and Brun, 2008). Even if mainly syn-orogenic, this detachment may have been active in the late stages of  
184 post-orogenic extension (Jolivet et al., 2010a).

185 In Italy, the Catena Costiera detachment is described as a mylonites and cataclasites complex, accounting  
186 for the post-orogenic exhumation of the blueschist-bearing metaclastics and marbles associated to the  
187 Lower Ophiolitic Unit (LOU, Jolivet and Brun, 2008; Rossetti et al., 2004) beneath the seemingly Upper

188 Ophiolite Unit, composed of an ophiolitic melange and carbonates (Bonardi et al., 2001). The Aspromonte  
189 detachment is a mylonite zone located on top of the Alpine blueschist Aspromonte unit, considered as a  
190 syn-to immediate post-orogenic structure (Heymes et al., 2010). Aspromonte and Stilo units on both sides  
191 of the detachment are both composed of para and orthogneiss.

192 Cretan detachments, described on top of the blueschist facies Phyllites-Quartzites nappe in northern  
193 Peloponnese (Jolivet et al., 2009b) and in Crete (Kilias et al., 1994) are represented by extensional  
194 mylonites and high-strain phyllites along the reactivated thrust at the base of the Gavrovo-Tripolitza  
195 nappe in Crete, or the underlying Tyros Beds in Peloponnese.

196 In the Rechnitz window group, the detachment zone is composed of serpentinites and quartz-micas  
197 mylonites (Cao et al., 2013) at the base of the Austro-Alpine greenschist facies Wechsel unit, and at the top  
198 of the blueschist-bearing Rechnitz Penninic unit. A limited HT imprint is recorded during the inferred  
199 post-orogenic reactivation of the Austro-Alpine basal thrust (Hoinkes et al., 1999).

200 All the detachment detailed above are therefore located within a tectonostratigraphic sequence with no  
201 significant systematic strength contrast, upper and lower plate being most of the time similar in  
202 lithologies, except the HP paragenesis partly preserved in lower plates. Most of the time, extensional  
203 strain localized on a former thrust. The reason why such thrusts are prone to extensional reactivation will  
204 be discussed further.

205

### 206 **2.1.3 Detachments at the base of a strong upper crustal unit**

207 Two close classes of detachment emerge from this review: some localize at the base of a strong upper  
208 plate and on top of a HP preserving cover unit and some at the base of a strong upper plate on top of a HT  
209 lower plate.

210 In Nigde massif, Turkey (NiD, table 1W) the amphibolite facies Gumusler formation migmatites (Whitney  
211 and Dilek, 1997) are exhumed beneath a massive ophiolite mostly composed of greenschist-facies gabbros  
212 (Goncuoglu et al., 1991). Mylonites and cataclasites developed prior to the Uçkapili leucogranite intrusion  
213 (Gautier et al., 2002) during an early extension stage affecting the Central Anatolian continental block.

214 In southern and eastern Rhodope (Bulgaria-Greece), the Kerdylion detachment and the Tokachka and  
215 Kechros detachment system (KerD and ToKeD, table 1M & N) developed at the base of high-grade rock  
216 units: the Vertiskos Gneiss complex (Brun and Sokoutis, 2007) and the Kimi complex garnet-bearing  
217 gneiss (Bonev and Beccaletto, 2007) respectively. They exhume the Southern Rhodope Core Complex, the  
218 Kesebir-Kardamos and Kechros complexes, all of them being represented by migmatites (Bonev et al.,  
219 2006) and amphibolite facies orthogneiss intruded by syn-kinematic granitoids (Vrondou and Symvolon  
220 granodiorites in Southern Rhodope, and the Papikjon granitoid in Eastern Rhodope). The base of upper  
221 plates is underlined by resistant lithologies such as the Therma and Volvi gabbros and basalts in Southern  
222 Rhodope or meta-ophiolitic lenses in Eastern Rhodope (Burg et al., 1996).

223 The Motajica Detachment, Croatia-Bosnia Hercegovina (MoD, table 1, Ustaszewski et al. (2010) is located  
224 at the base of the Tisza-Dacia unit ophiolite and at the top of an amphibolite facies meta-accretionary  
225 wedge complex intruded by the Motajica granite a few Myr prior to extension initiation (Pamić et al.,  
226 2012).

227 In Kazdag, Turkey, extension localized along the northern Alakeçi and southern Selale detachments  
228 (AlSeD, table 1S). Mylonites, metaserpentinites and breccias separate the upper plate Cetmi ophiolitic  
229 melange from an amphibolite facies basement unit (Beccaletto and Steiner, 2005; Okay and Satir, 2000)  
230 topped with migmatites. The Cetmi, Karakaya and Denizgoren ophiolitic units all rest at the base of the  
231 upper plate, suggesting a possible reactivated obduction thrust as a precursor for the extensional  
232 detachments.

233 These detachment therefore superimpose resistant lithologies on top of HT metamorphic rocks, some  
234 other superimpose the same type of resistant units on top of weaker HP-bearing cover units as reviewed  
235 in the following.

236 In the Alpi Apuane, northern Italy, the Calcare Cavernoso mylonites (CalCa, table 1G, (Carmignani and  
237 Kligfield, 1990) are found between the upper Tuscan nappes made out of massive carbonates at the base  
238 and the subjacent blueschist-facies phyllites and metasandstones of the Massa and Autochthonous units.  
239 Localization of strain is due to intense cataclasis of anhydrite and dolostone along a reactivated thrust.  
240 The Liguride ophiolite on top of the whole lithotectonic pile has been preserved from most of the  
241 extensional tectonics.

242 In Corsica, the Tenda Shear Zone and the Balagne-Nebbio unit basal detachment constitute a multi-level  
243 detachment system (Jolivet et al., 1990; Marroni and Pandolfi, 2003) related to post-orogenic extension. In  
244 the Cap Corse antiform, the non-metamorphic ophiolitic Balagne nappe directly rests on the Schistes  
245 Lustrés calcschists that preserved eclogite to blueschist facies assemblages. On the eastern edge of the  
246 Tenda massif, ophiolitic units and basement slices associated to the Schistes Lustrés are separated from  
247 the blueschist facies metagranitoids by a thick and complex mylonite zone developed to the expense of the  
248 latest.

249 All these examples therefore lie in intermediate position between the HT end-member or the HP  
250 preserving end-member and a theoretical third end-member, which would be represented by a  
251 detachment localized at the base of an anomalously strong upper plate.

252

253 Some other Mediterranean detachments also lie in intermediate position in the three end-member  
254 classification proposed here (Figure 5).

#### 255 **2.1.4 Detachments with intermediate characteristics**

256 The Malaguides-Alpujarrides detachment (MAD, table 1A, Lonergan and Platt, 1995) in the Betics and the  
257 Naxos-Paros detachment (NaPaD, table 1Q, Gautier et al., 1993) in Central Cyclades plot along the HT and  
258 HP preserving members joint (Figure 5). Both indeed localized at the top of a HP bearing unit, and  
259 developed during a thermal reequilibration stage. This thermal reequilibration is more intense in the  
260 Naxos-Paros lower plate where prealpine basement underwent pervasive anatexis during exhumation  
261 (Vanderhaeghe, 2004), than in the Alpujarrides units, where high-pressure amphibolite facies overprint  
262 induced local migmatization in the top Herradura unit and pervasive resetting of radiochronometers  
263 (Azañón and Crespo-Blanc, 2000).

264

265 Other complex detachment systems such as the North Cycladic Detachment System in the Aegean (NCDS,  
266 table 10, Jolivet et al., 2010a) or the Menderes detachments in Turkey also plot within the theoretical  
267 triangular classification, but show a singular evolution through time as detailed below.

## 268 **2.2 Evolution of detachments through time**

269 Three features identified in this review allow discussing how detachment systems  
270 evolve through time. When present, the nature of the collapse basin associated to  
271 extension along the detachment has been systematically noticed (Figure 6) and evidence  
272 how topography locally evolved during extension. The Cyclades and Menderes exhibit  
273 complex cases of multi-level detachment systems that evolved through time. Eventually  
274 relationship between ductile strain and brittle strain gives indices of how detachment  
275 systems behave when crossing the brittle-ductile transition.

276

### 277 **2.2.1 Evolution of basins**

278 Most of the syn-tectonic basins described in the hanging-walls of detachment systems  
279 have marine affinities or evolved through time from continental to shallow marine,  
280 suggesting a topography close to sea level, and most of the time subsiding during  
281 extension (Figure 6 and table 1, line 2.1). Accommodation is most of the time moderate,  
282 with shallow marine coarse facies, but turbidites are described in the Kazdag massif,  
283 suggesting a deep depositional environment (Bonev and Beccaletto, 2007). The opposite  
284 trend from marine to continental is only described on Naxos, where marine sandstones  
285 and pelites evolve into fluvial deposits (Bargnesi et al., 2013; Kuhlemann et al., 2004),  
286 implying that accommodation is decreasing with time, unless regional parameters  
287 actually prevail on local processes . Even in the case of detachments initiated in syn-  
288 orogenic conditions, such as in Crete, where possibly lacustrine breccias are described at  
289 the base of Neogene basins (Jolivet et al., 1996), no proper intra-montane collapse  
290 basins developed during activation of detachments in the Mediterranean realm. This

291 implies that initial topography was close to or rapidly lowered to sea-level while MCC  
292 developed.

293

### 294 **2.2.2 Evolution of strain localization**

295 Complex detachment systems evolution can be traced in the three end-member  
296 classification proposed here. As for the Cyclades, the north-verging North Cycladic  
297 Detachment System (table 10, Jolivet et al., 2010a) and south-verging Western Cycladic  
298 Detachment System (table 1R, Grasemann et al., 2011) are composed of several mylonite  
299 zones that successively localized strain during regional extension. In Northern Cyclades,  
300 the Tinos Detachment first localized along a former thrust at the top of the Cycladic  
301 Blueschist Unit and beneath the Upper Cycladic Nappe, constituted of a thick and  
302 resistant greenschist facies ophiolitic complex . Strain sequentially localized upward on  
303 the Livada detachment and then Mykonos detachment, while granites progressively  
304 intruded the lower plate and thermal reequilibration of the nappe pile occurred. On  
305 Mykonos, syn-tectonic granite directly roots in the highly strained migmatites of the  
306 lower plate (Denèle et al., 2011), evidencing a sharp contrast in thermal history between  
307 the upper and lower plates. In Ikaria, the Fanari ductile to brittle detachment located at  
308 the top of the syn-kinematic I type Raches granite and the ductile Agios Kyrkos shear  
309 zone show strain and age patterns similar to the North Cycladic Detachment System, and  
310 can be considered as the eastward extension of the same structure (Beaudoin et al.,  
311 2015; Laurent et al., 2015). The North Cycladic Detachment System therefore evolves  
312 from lower bound toward the HT end-member in the ternary classification proposed,  
313 with a perennial detachment zone relocalizing stepwise at the base of the upper plate  
314 while the lower plate exhumes. The Western Cycladic Detachment System described on

315 Kea, Kythnos and Serifos islands possibly followed the same evolution with initial  
316 detachment located at the base of the Pelagonian serpentinites and schists and  
317 progressively moving upward in the nappe pile as the thermal effect of syn-tectonic  
318 intrusions (on Serifos and Lavrion) modified the crustal rheological properties  
319 (Rabillard et al., 2015).

320 In Menderes, the initial structure that accommodated extension during Hellenic slab  
321 retreat is the Simav Detachment (SiD, table 1T, Isik and Tekeli, 2001). This large scale  
322 high strain zone is localized at the base of the Izmir - Ankara Ophiolite, within a HP  
323 nappes complex preserving eclogites and blueschists (Afyon, Tavsanli and Oren nappes  
324 (Plunder et al., 2013). Rocks in its footwall mainly consist of amphibolite facies gneiss  
325 with the metasedimentary Selimiye unit in uppermost position. Early northern  
326 Menderes granites are considered as syntectonic to SiD (Isik and Tekeli, 2001), due to  
327 development of high temperature foliation in these in the vicinity of SiD and overlap  
328 between their U-Pb zircon crystallization ages and cooling ages of biotites within  
329 derived mylonites (Dilek et al., 2009) in the early miocene . The younger Alasehir and  
330 Kucuk-Menderes Detachment systems, Central Menderes, Turkey (AlKuD, table 1T),  
331 subsequently cut across Menderes and divide it in Northern, Central and Southern  
332 domains. Rocks beneath and above the Alasehir and Buyuk Menderes detachments  
333 therefore derive from the same amphibolite facies metamorphic sequence (Bozkurt and  
334 Oberhänsli, 2001). The north-verging Alasehir detachment located at the top of the  
335 Salihli syn-tectonic granodiorite (Isik et al., 2003) in the late Miocene (Dilek et al., 2009)  
336 and the antithetic detachment at the base of the Buyuk Menderes graben also relates to  
337 the late Miocene intrusion of Menderes granites (Dilek et al., 2009). Even if recent  
338 migmatites are actually present within Menderes (Bozkurt, pers. comm.) the sole  
339 localization mechanism proposed so far for these detachments is the weakening due to

340 granite intrusion. Strain indeed localized at the top of these plutons that evolve from  
341 isotropic in their core toward mylonites in the vicinity of detachment (Dilek et al., 2009).  
342 Mylonitic foliation is marked by biotites within granites and greenschist facies  
343 assemblages within mylonites. Strain therefore progressively localized during granite  
344 cooling and exhumation. The central detachments accommodated the late stages of  
345 intra-Menderes extension in middle miocene times, while the top Menderes northern  
346 Simav detachment progressively ceased localizing strain (van Hinsbergen, 2010).  
347 These complex long-life detachment systems systematically show an initial localization  
348 along intra-crustal rheological contrasts inherited from nappe stacking stages, the base  
349 of ophiolites and HP-bearing cover units being prone to such localization. The later  
350 stages are associated to HT metamorphism and/or syntectonic granite intrusions. Strain  
351 is then localized higher up in the tectonic pile (as in the Cyclades) on a structure  
352 synthetic to deeper precursors or symmetrically neofomed in the core of the already  
353 exhumed lower plate (as in Menderes).

354

### 355 **2.2.3 Relationships with brittle deformation**

356 When evidenced, relationships between brittle strain in the upper and lower plates and  
357 the ductile to brittle strain along the detachment system were systematically noticed  
358 (Table 1, items 2.4, and 4.3). In most cases (Catena Costiera, Alpi Apuane, Rechnitz,  
359 Northern Cyclades, Northern Menderes) brittle strain is expressed as high- to low-angle  
360 normal faults, developed in the upper plate and rooting in the detachment zone. More  
361 rarely brittle deformation cuts across and offsets the main detachment system. When  
362 expressed as late stage high-angle normal faults (as in Balagne, Corsica, in the Motajica



363 window, in southern and eastern Rhodope or Crete), this brittle strain can be viewed as  
364 late features accommodating the late stages of extension or unrelated later tectonics.  
365 The case of Elba island, with the development of the Zuccale fault, is a rare described  
366 case of a low angle brittle normal fault localizing independently of ductile inheritance  
367 and cutting across the extended nappe pile (Collettini and Holdsworth, 2004). This  
368 feature is also recognized today along the Apennine detachment system, with the low  
369 angle normal active Alto Tiberina fault cross-cutting all the thrust structures (Chiaraluce  
370 et al., 2007), which system the Zuccale fault zone also pertains to **(Collettini et al.,**  
371 **2006)**.

372 Interpretation of relationships between ductile and brittle strain is not unique:  
373 depending on whether faults bounding basins are rooted on or cut across the  
374 detachment system, Menderes basin system can be viewed as syn-tectonic collapse  
375 structures (according to VanHinsbergen, 2010 for instance) or a wide rift zone with  
376 periodic rift basins obliterating the core complex structure (Ring et al., 2003).

377

### 378 **2.3 Rheological implications of model end-members**

379 The three end-members proposed in this review are not equally straightforward in  
380 terms of rheological implications. The HT end-member actually represents the most  
381 widely accepted definition of MCC, with a hot and hence low-viscosity lower crust. It has  
382 been quantitatively considered in many thermo-mechanical models (e.g. Rey et al., 2009;  
383 Tirel et al., 2006). A theoretical HS end-member with resistant upper crust, constituted  
384 of massive mafic or ultramafic lithologies thrust on top of a less resistant crust, can also  
385 be considered. Intra-crustal rheological contrast is here produced by the tectonic

386 inversion of the rheological profile (Huet et al., 2010) and has been tested via thermo-  
387 mechanical modelling. The third HP end-member, representing detachments developing  
388 at the top or within HP cover units is less straightforward in terms of strength profile.  
389 Why would strain preferentially localize in or along such units in the post-thickening  
390 nappe pile ? (Huet et al., 2010) considered the Cycladic Blueschist Unit as a weak layer  
391 intercalated between the more resistant Upper Cycladic Unit ophiolite and the Cycladic  
392 Basement Unit constituted of granitoids mainly. Before testing the effect of such a  
393 layering, its relevance must be discussed. HP cover nappes can be considered weak due  
394 to the nature of their protolith. Initially constituted of shales and carbonates, they are  
395 fine grained, their water content is high and they are strongly foliated, thus being prone  
396 to localize deformation. Nevertheless, metamorphic transformations may significantly  
397 change their mechanical behaviour. In order to quantitatively discuss this point we  
398 performed thermochemical modelling of the impact of exhumation on average crustal  
399 lithologies (Figure 7). Modal evolution of a metabasalt, a metagranite and a metapelite  
400 with 10 wt% added water (dry compositions basalt and granite : Le Maitre (1976), shale  
401 : Boggs 1995) have been modeled with Theriak-Domino suite (de Capitani and  
402 Petrakakis, 2010) along different pressure-temperature paths from blueschist (1.2 GPa,  
403 500°C) to lower greenschist facies (0.3 GPa, 200°C) using Holland & Powell database  
404 (Holland and Powell (2004) with Diener et al. (2007) activity model for amphiboles).  
405 Two cases can be considered : a path with thermal reequilibration coeval with  
406 decompression (path 1, Figure 7), and a path with decompression in the greenschist  
407 facies prior to substantial cooling (Path 2, Figure 7), more representative of the PT paths  
408 documented in HP units exhumed in Mediterranean MCC (Figure 3). In the case of a  
409 basalt, exhumation is marked by decrease in amphibole amount (due to glaucophane  
410 destabilization) whatever the path. Epidote is retrogressed into lawsonite along path 1

411 only. No substantial change in phyllosilicate content is observed whatever the  
412 considered path. In the case of a granite, the only substantial modal change modeled is a  
413 decrease then increase in phyllosilicate content along path 2. Eventually, in the case of a  
414 metapelite, epidote-lawsonite transition is also observed, and phyllosilicate content  
415 rises close to 50 vol% along path 2. Considering the amount of phyllosilicates as one of  
416 the first order mineralogical control on rocks strength in the blueschist/greenschist  
417 temperature range (Gueydan, 2004), the comparison of its evolution along paths for the  
418 different lithologies can yield the evolution of strength contrasts within a composite  
419 nappe pile (Figure 7D&H). As expected metapelites exhibit the highest phyllosilicates  
420 contents. No substantial contrast appear along path 1, while the more realistic two-stage  
421 PT path 2 exhibits a strong contrast between metapelites, reaching 40 vol. %  
422 phyllosilicates and other lithologies between 500 and 400°C i.e. within the ductile strain  
423 domain. It seems therefore that initially weak metasedimentary sequences are  
424 furthermore weakened by retrogression in the greenschist facies. Fluid transfers  
425 implied by these transformations also can lead to strength changes during exhumation.

426 HP cover units can therefore be considered at first order as weaker zones interlayered  
427 in the nappe pile, and therefore prone to localize deformation during syn-to post-  
428 orogenic extension. It seems from the presented review that the development of a  
429 perennial detachment during extension of a beforehand-thickened crust is therefore  
430 possible when a critical intra-crustal strength contrast appears, due to thermal  
431 reequilibration (HT MCC case) or due to inheritance (strong upper crust and weak  
432 middle crust). This conceptual model, can be quantitatively explored with thermo-  
433 mechanical models in order to estimate the order of magnitude of the required intra-  
434 crustal strength contrast and to assess how the different end-members can yield a  
435 detachment structure.

436

### 437 **3. Numerical modelling**

#### 438 **3.1 Numerical code used**

439 The numerical code used in the present study is based on the FLAC algorithm (Cundall,  
440 1989) and its subsequent evolutions (Burov and Poliakov, 2001; Poliakov et al., 1993;  
441 Yamato et al., 2007). Conservation of momentum equation and heat equation are  
442 iteratively solved. No radiogenic internal heating was implemented in the present  
443 simulations, mainly due to uncertainties on its value in the case of a crust resulting from  
444 complex nappe stacking. Thermomechanical coupling is enforced using Boussinesq  
445 approximation for the computation of density, including thermal stresses. Temperature  
446 is advected with the mesh within the Lagrangian formulation of the code. The rock  
447 behaviour is approximated by explicit visco-elasto-plastic rheology. Ductile deformation  
448 is modeled with a power-law and brittle deformation with a Mohr-Coulomb plastic flow  
449 law (table 3). The effective rheological behaviour is determined by current strain-rate,  
450 state of stress and temperature (Le Pourhiet et al., 2004). A marker-based remeshing  
451 procedure allows for handling of very large strains and displacements (Yamato et al.,  
452 2007). To focus on the influence of the rheological stratification, as in (Huet et al., 2010),  
453 the model setups have been deliberately simplified. Erosion, shear-heating, partial  
454 melting and mineral phase transitions are not considered in the computation.

455

#### 456 **3.1 Initial set-up and initial boundary conditions**

457 Experimental design (Figure 8) directly derives from Huet et al. (2010). Thermal  
458 boundary conditions are 0°C at the surface and fixed initial temperatures at the bottom

459 (1000, 1200 and 1400 °C at 90 km, implying initial Moho temperatures between 398  
460 and 1015°C, as listed in table 4). Lateral heat flow is set to zero. These thermal  
461 conditions allow to cover the complete crustal thermal gradient range (dashed curves  
462 on Figure 3) from low (13°C/km), as inferred from PT paths in the Cyclades for instance  
463 (Figure 3 and Huet et al., 2010) to moderate (18°C/km) as explored in previous thermo-  
464 mechanical modelling studies (Tirel et al., 2008 for instance).

465 Both sides of the models are assigned free slip condition. As in Huet et al., 2010,  
466 asymmetric lateral velocity is applied with 1 cm/yr on the lefthand side and 0 cm/yr on  
467 the righthand side.. Initial stretching rate for a 210 km long crustal segment is therefore  
468  $2 \cdot 10^{-15} \text{ s}^{-1}$ . This velocity boundary conditions corresponds to a mean value in the  
469 sensitivity tests to strain rates in numerical studies (Tirel et al., 2008), or in the velocity  
470 range used for analytical modeling (Buck, 1990). Upper surface is free, and bottom  
471 behaves as a Winkler foundation, ensuring local isostatic compensation (Burov and  
472 Poliakov, 2001). The initial velocity field is linearly interpolated from the boundary  
473 conditions and the stress is initially lithostatic and isotropic. No resistance or density  
474 anomaly is prescribed to localize strain, a random noise of 5 MPa has been instead  
475 added to the mean cohesion value (20 MPa, table 3) in the upper crust, except on a 20  
476 km wide zone on the left hand side of the model where cohesion is set to 30 MPa in  
477 order to prevent artificial strain localization on the fast moving boundary (Huet et al.,  
478 2010).

479 The model is initially composed of a 210 km wide and 30, 45 or 60 km thick two-layers  
480 continental crust resting on top of lithospheric mantle. Initial mesh grid size is 0.75 x  
481 1.25 km. All crustal layers have the same nominal density and thermal expansion  
482 coefficient (table 3), so that only temperature contrasts can yield density contrasts.

483 Mantle properties are the same in all simulations, its viscous behaviour is modelled as  
484 the one of dry dunite (Chopra and Paterson, 1984). Viscosity parameters for crustal  
485 layers derive from experimental data from quartz-diorite (Ranalli and Murphy (1987) as  
486 used in Tirel et al., 2008), dry quartzite (Ranalli and Murphy (1987) as used in Huet et  
487 al., 2011) and wet quartzite (Kirby and Kronenberg, 1987). Crustal structures can be  
488 rheologically homogeneous (27 simulations: d/d, q/q and w/w series in table 4) or  
489 heterogeneous. Among heterogeneous crust simulations, 18 were performed with upper  
490 and lower crust the same thickness (evenly layered crust, table 4). 6 simulations were  
491 performed with total crustal thickness 45 km composed with quartz-diorite upper crust  
492 and dry quartzite lower crust, and 15 or 30 km thick upper crust.

493

### 494 **3.2 Results**

495 The 51 simulations run in this study are compared on the basis of the strain geometries  
496 produced by extension as well as P-T paths followed by the possibly exhumed lower  
497 crust. The different resulting geometries, exemplified in Figures 9, 10 and 11 are briefly  
498 described before a more statistical approach is proposed for all simulations. Only 3  
499 cases out of 51 (X in table 4) yielded numerically unstable computation, with mantle  
500 rupturing from the very first stretching steps. These simulations, performed with cold  
501 conditions (bottom temperature 1000°C) are considered as geologically meaningless.  
502 Narrow rifts simulations all exhibit localized strain zones in the upper crust and upper  
503 mantle that connect across the ductile lower crust to stretch the complete crust to  
504 complete disruption after less than 10 Myr extension. They are characterized by high  
505 amplitude Moho deflection and limited lower crustal flow. Margin geometries produced  
506 exhibit complex geometries due to simple or complex rift systems (Figure 9C) isolating

507 upper crustal blocks. Wide rift geometries (Figure 9B, D and F) show no transcrustal  
508 shear zones developing, upper crust disruption is compensated by lower crustal flow  
509 and Moho deflection is limited. Nevertheless, as analytically predicted by Buck (1991),  
510 lower crust resistance to flow is higher than plastic yield strength of the upper crust and  
511 strain relocalizes laterally, before an extended lower crust dome is formed. Strain  
512 localization shifts laterally in time with a wavelength determined by initial set-up.  
513 Metamorphic core complex geometries can be sub-divided into two sub-types :  
514 asymmetrical domes with a perennial detachment system assisting exhumation of the  
515 lower crust over the whole stretching event (Figure 10), and more symmetrical  
516 geometries with lower crust exhumation and strain localizing between two sub-domes  
517 (double-dome core complexes, Figure 11, DDCC in table 4). In both cases, the upper  
518 crust disruption is compensated by lateral and upward flow of the lower crust into a  
519 single perennial core. In the perennial detachment metamorphic core complex case  
520 (Figure 11, PDCC in table 4), strain relocalizes at the upper lower crust boundary  
521 yielding an asymmetric flow pattern in the lower crust and the persistence of a  
522 detachment system at the top of the dome structure with younging upward strain ages .  
523 The youngest rocks exhumed are then found beneath this detachment system, and  
524 exhibit "cool" PT path with no heating associated to decompression (Figure 10 B, F & G).  
525 In the double-dome core complex case (Figure 11) the high strain zone is located at the  
526 core of the exhumed lower crust, implying a symmetric flow pattern and exhumation  
527 ages younging from the edges to the core (Figure 11B & F). The later rocks exhume, the  
528 higher the amplitude of their decompressional heating (Figure 11G). Rocks exhumed  
529 within or close to detachments systematically show decompression paths along lower  
530 thermal gradients (Figures 10G & 11G). This remarkable feature is due to progressive  
531 cooling by the overriding upper unit, and might be less pronounced with viscous heating

532 implemented in the code used. Thermal impact of viscous heating along extensional  
533 detachment systems has been estimated about 50°C temperature excess during  
534 decompression of hanging-wall units (Souche et al., 2013), and would therefore affect  
535 PT path shapes only on the second order.

536 Resulting geometries occur logically according to initial set-up (Figure 12). Narrow rifts  
537 are produced by thin crust simulations whatever their thermal state, except when lower  
538 crust is drastically weak. Wide rift geometry is indeed produced for wet quartzite lower  
539 crust simulation with initial crustal thickness 30 km and initial Moho temperature above  
540 400°C (Figure 12C). On the other hand double-dome core complexes geometries are  
541 produced in thick crusts, whatever their thermal state, except when especially strong.  
542 Simulation with 60 km thick quartz-diorite homogeneous crust and  $T_{\text{Moho}} = 752^\circ\text{C}$   
543 indeed yield a wide rift pattern. The most sensible crustal thickness is 45 km. With this  
544 thickness, weakening of the whole crust promotes a switch from wide rift to double  
545 dome MCC pattern (Figure 12 A, B & C). Weakening of the lower crust only has the same  
546 effect but intermediate strength lower crust simulations (Figure 12D) yield wide-rift  
547 geometry when cold and perennial detachment MCCs pattern when warmer. The  
548 position of the intra-crustal strength contrast within the crust also has an impact.  
549 Thinner lower crust geometries (line 2/3 on Figure 12F) yield a narrow rift pattern  
550 when cold, which can be explained by an easier connection between localized strain in  
551 the upper crust and in the mantle and an overall less ductile rheology (higher  
552 ductile/brittle ratio). Thicker lower crust simulations (lower ductile/brittle ratio, line  
553 1/3 on Figure 12F) yield more distributed extension mode (wide rift or double dome  
554 MCCs according to thermal state). Perennial detachment MCCs are produced for  
555 intermediate and maximal initial Moho temperatures and low to intermediate lower  
556 crust thicknesses. From this analysis, several general comments arise: i- thermal state is



557 not the dominant controlling factor on extension modes, at least in the range explored  
558 here, ii- crustal thickness has a critical impact in the 30-60 km range, iii- perennial  
559 detachment MCCs are developed only in specific conditions with intermediate initial  
560 crustal thicknesses and intermediate intra-crustal strength contrasts.

561 In order to explain them with unified criteria, occurrences of the four final geometries  
562 described here (narrow and wide rifts and metamorphic cores with perennial  
563 detachment or double dome geometry) have systematically been reported against initial  
564 thermal and rheological parameters for the tested lithospheres (Table 4 and Figure 13).  
565 Intra-crustal strength contrast, i.e. ratio between the strength at the brittle-ductile  
566 transition depth and viscous strength of the crust at Moho depth as well as the  
567 integrated lithospheric strength expressed in MPa.km, have been calculated from the  
568 initial conditions of all models. These parameters have been chosen for their relevance,  
569 as well as for their easy calculation for already published comparable studies (cf  
570 discussion). Tested integrated strengths range over more than 1 order of magnitude,  
571 cold and thin crusts leading as expected to stronger lithospheres than warm and thick  
572 ones. Intra-crustal strength contrasts range over 4 orders of magnitudes, with the thin  
573 and cold crusts having contrast ratios lower than 10 and warm and thick crust  
574 exhibiting ratios higher than  $10^4$ . Rift geometries are produced for high lithospheric  
575 strength and low strength ratios. Only such lithosphere can indeed reach their plastic  
576 yield strength and connect localized upper crust strain to localized strain in the mantle.  
577 Wide rift geometries are produced over a wide range of integrated strength but a  
578 narrow range of strength ratio. This reflects the close balance between upper crust  
579 strength and lower crust ability to flow implied by the development of wide rifts. Double  
580 dome MCCs are develop over a wide range of integrated strengths for strength ratios  
581 higher than  $10^3$ . The specific conditions prone to the development of perennial

582 detachment appear as a narrow intra-crustal strength contrast range close to  $10^3$ . In this  
583 range, all the possible extension modes are actually produced; only rifts are produced  
584 for contrasts lower than  $10^{2.5}$  ( i.e.  $\sim 300$ ) and only double dome MCCs are produced for  
585 contrasts higher the  $10^{3.5}$  ( i.e.  $\sim 3000$ ).

586 The effect of each tested parameter can be precised with series having one varied  
587 parameter only (dashed lines on Figure 13). As expected, the initial crustal thickness  
588 and initial thermal state impact both parameters. Thick and/or warm crusts yield high  
589 intra-crustal strength contrast for low integrated strength. Layering has an impact on  
590 contrast more than on integrated strength. The intra-crustal strength contrast  
591 parameter is actually irrelevant to investigate the effect of the depth of inherited  
592 strength discontinuity within the crust (crosses on Figure 13). More sophisticated  
593 parameters would help, but would necessarily be more design-dependent.

594 The parametric study presented here yields some qualitative and quantitative  
595 conclusions about the boundary conditions and lithosphere properties prone to the  
596 development of MCC patterns. First, MCCs can form in moderately thickened crust along  
597 geotherms as low as  $13^\circ\text{C}/\text{km}$ . Depending mainly on the intrinsic strength ratio between  
598 the upper and lower crust, two types of MCCs can develop: double domes for crusts with  
599 a high strength ratio, MCCs with a perennial detachment at their top for crusts with a  
600 strength ration close to  $10^3$ . Exhumed lower crust material shows limited heating during  
601 decompression in MCCs with a perennial detachment, and only material exhumed late in  
602 double-domes exhibits substantial heating. MCC pattern development is therefore  
603 thermally compatible with the preservation of HP-LT paragenesis in their core,  
604 especially on their limb.

605

## 606 **4. Discussion**

### 607 **4.1 Parameter study and published MCC models : critical intra-crustal strength** 608 **contrast for MCC formation**

609 In order to test our parameter study against published analogue studies, the same  
610 parameters (intra-crustal strength contrast and integrated lithospheric strength) were  
611 computed for published simulations with comparable initial set-up (Figure 14). 32  
612 simulations from Tirel et al. (2008) and 9 cases study in Huet et al (2011) were  
613 performed with the same numerical code as present and layered or homogeneous  
614 crustal geometries. Among the 4 simulations presented in Rey et al. (2009) and  
615 performed with the Ellipsis finite element code (Moresi et al., 2003), the 2 melt-free  
616 cases with varied stretching rate ( $2 \cdot 10^{-16}$  and  $2 \cdot 10^{-15} \text{ s}^{-1}$ ) can yield representative initial  
617 crustal strength contrast and integrated strength values. In the study performed by  
618 Schenker et al. (2012), with the I2VIS code (Gerya and Yuen, 2003) 12 initial geometries  
619 were tested with a rheologically homogeneous crust and varied initial thickness and  
620 Moho temperatures. In these two studies, partial melting is implemented as a viscosity  
621 drop down to a threshold value ( $10^{18} \text{ Pa.s}$  in Rey et al., 2009,  $10^{17} \text{ Pa.s}$  in Schenker et al.,  
622 2012). Eventually, the two-layer wedge shape crust simulations without prescribed  
623 weak zone (A and B type models) from Wu et al. (2015), performed with a modified  
624 version of PARAVOZ (Tan et al., 2012) can be plotted on the same diagram as segments  
625 between their thin and thick ends. Compilation of these simulations with ours give a 108  
626 simulations dataset, produced by different numerical codes, different crust and mantle  
627 representative materials, and a variety of boundary conditions. They all plot along the  
628 same trend in the intra-crustal strength contrast vs integrated lithosphere strength.  
629 Tirel et al. (2008) explored a linear domain across the whole possible field, while Huet et

630 al. (2011) and Rey et al. (2009) focused on weak lithospheres and intermediate strength  
631 profiles. The equivalence between the present-study classification and other studies is  
632 not straightforward. The "double-domes" produced by Rey et al. (2009), the spreading  
633 domes in Huet et al., (2011) and the double-dome MCCs in the present study can be  
634 considered as equivalents. The "true rifts", "wide rifts" and "MCCs" produced in Huet et  
635 al. (2011) are respectively synonyms in terms of dynamics with rifts, wide rifts and  
636 perennial detachment MCCs produced here. Even if Schenker et al. (2012) mainly  
637 focused on the impact of the depth of the heat source (within the crust or within the  
638 mantle) on the lithospheric response, their classification also has a morphological basis.  
639 The "dysharmonic dome" and "lower crustal dome" types distinguished actually  
640 respectively correspond to the perennial detachment mode and the double-dome core  
641 complex types used here. "Gneiss domes" and "narrow rifts" structures produced in Wu  
642 et al. (2015) with 2 layer crusts show some periodicity that could relate them to wide  
643 rifts produced in the present study. In first approximation, "narrow rifts" and  
644 "continental margins" types can be considered as localized extension modes with crustal  
645 necking, while the "gneiss dome" mode compares with the double-dome MCCs produced  
646 here. In Tirel et al. (2008) the only available categorization is "crustal necking" vs "MCC".  
647 Even if presented examples of MCCs rather exhibit perennial detachment core  
648 complexes geometries, we keep the classification as simple as published. This  
649 compilation eventually shows that rifts or MCCs can be produced over a wide range of  
650 integrated lithospheric strength. Intra-crustal strength contrast seems a sensitive  
651 parameter for extension mode prediction: here again, only rifts are produced for  
652 contrasts lower than  $10^{2.5}$  ( i.e.  $\sim 300$ ) and only double dome MCCs are produced for  
653 contrasts higher the  $10^{3.5}$  ( i.e.  $\sim 3000$ ).

654 Tirel et al. (2008) extensively explored the effect of strain rate on extension mode, and  
655 Rey et al. (2009) also varied this parameter. Their results on this diagram show that an  
656 increase of 1 (for Rey et al., 2009) or close to 1 (for Tirel et al., 2008) order of magnitude  
657 only slightly modifies the crustal properties, and only changes extension mode for  
658 lithospheres in the critical domain next to  $10^3$ . Wu et al. (2015) who studied the effect of  
659 layered wedge shape come to the same conclusion as here: intrinsic strength contrast  
660 due to inherited layering is a first order parameter for changing extension mode. The  
661 realistic values used in that study actually cover a 2 orders of magnitude range in  
662 strength contrast between the strongest (plagioclase above dry quartz) and the weakest  
663 (plagioclase above wet quartz with 0.12 % added water).

664 Scaling of analogue models in Figure 14 diagram is impossible, but viscosity and strain  
665 rates values used in (Brun et al., 1994) allow the calculation of a  $10^{1.2}$  intra-crustal  
666 strength contrast in the tilted-block mode extension and  $10^{2.39}$  contrast in MCC mode.  
667 Here again, an increase about 1 order or magnitude is sufficient to promote MCC pattern  
668 development.

669 Since the first order parameter often considered as critical for MCC pattern development  
670 is initial thermal state (Buck, 1991, Tirel et al., 2008) to such an extent that MCCs are  
671 sometimes considered as restricted to warm lithospheres, the effect of intra-crustal  
672 strength contrast can be evaluated in regard to initial Moho temperature (Figure 15). All  
673 simulations considered here cover a wide domain of initial thermal conditions between  
674  $400^{\circ}\text{C}$  and  $1340^{\circ}\text{C}$  Moho temperatures, i.e. much wider than the expected possible  
675 values, even in the most severe post-orogenic thermal relaxation. The present study  
676 produces MCCs with initial Moho temperature as cold as  $680^{\circ}\text{C}$ . Huet et al. (2009) and  
677 Wu et al. (2015) yielded the same result, highlighting that cold MCCs constitute a

678 thermo-mechanically sound concept. The rift patterns with the initially warmest  
679 lithosphere were actually produced by Tirel et al. (2008). The domain in which the  
680 different models can give any of the distinguished extension modes actually span from  
681 470°C to 950°C initial Moho temperature, meaning that thermal state, in its admitted  
682 range, is not a critical parameter for predicting lithospheric extension mode.

683

## 684 **4.2 Numerical insights versus Mediterranean examples**

### 685 **4.2.1 Natural vs numerical categorization**

686 The parametric study presented here, together with a compilation of similar numerical  
687 modelling of continental extension modes leads on one hand to a 4 end-member model,  
688 with double domes MCCs, perennial detachment MCCs, wide rift and narrow rift as  
689 possible resulting geometries. On the other hand, the review of Mediterranean  
690 extensional MCCs points to a 3 end-member categorization of MCCs : HT MCCs with a  
691 high temperature lower unit, HP MCCs with an interlayered high-pressure nappe  
692 preserved in the vicinity of the detachment, and High Strength (HS) upper crust MCCs  
693 localizing strain at its base. The juxtaposition of both reviews enables discussing the  
694 mechanical properties behind the three natural types defined here. Actually each of the  
695 numerical studies also explored a ternary diagram with 3 end-members : high  
696 temperature, weak middle crust and strong upper crust (Figure 15 inset) that resembles  
697 the ternary diagram used for classification of natural Mediterranean cases. Weak middle  
698 crust relate to the HP natural end-member, and strong upper crust to HS end-member.  
699 Tirel et al. (2008), Schenker et al. (2012) as well as the high temperature experiments  
700 conducted here confirm the acknowledged behaviour of continental lithosphere with

701 high thermal gradient (HT end-member). Within this category the distinction of double-  
702 dome vs perennial detachment MCCS will be discussed further. Huet et al. (2009)  
703 explored the effect of an inter-layered weak zone, representing a HP metasedimentary  
704 nappe in natural cases. In this respect, they explored the left joint of the ternary  
705 diagram. The present study, with 2-layer crust lithospheres, as well as Wu et al. (2015)  
706 *pro parte* explored the right-hand joint of the diagram. Numerical modelling therefore  
707 evidences the mechanical feasibility of all natural end-members.

708

#### 709 **4.2.2 Natural double domes and perennial detachment MCCs in nature**

710 Not all natural cases can be plotted in the diagram derived from numerical simulations  
711 (Figure 16), since their initiation thermal conditions and intra-crustal strength contrasts  
712 remain uneasy to assess. Nevertheless, some of them can be qualitatively localized and  
713 tend to confirm the relevance of this categorization. The description of double-domes by  
714 Rey et al. (2009) is based on a two dimensional projection of Naxos core complex main  
715 features. Schenker et al. (2012) also consider Naxos as a natural case for their collisional  
716 heat induced migmatitic core complex, while Rhodope represents the asthenospheric  
717 heat induced migmatitic core complex. These cases (table 1) all illustrate the high  
718 temperature-high strength(Figure 16) end of our simulations, with low-viscosity  
719 migmatites exhumed in the core of the lower plate beneath a detachment located at the  
720 base of an unmetamorphosed ophiolite-bearing unit. Among cold MCCs, the Catena  
721 Costiera detachment (Rossetti et al., 2004) exhumed blueschist continental metaclastics  
722 and marbles beneath a greenschist ophiolitic melange. Mylonites and cataclasites  
723 derived from weak marbles and metapelites, implying a strong inheritance effect. Strain  
724 indicators show a pervasive asymmetric sense of shear toward NNW, related to the

725 activation of a single detachment zone. Other large MCCs such as Northern Cyclades,  
726 Motajica, Menderes or Betic cordilleras actually show evolution through time that can  
727 relate to the evolution of the mechanical properties of the crust they develop in.

728 Compilation of PT paths in metamorphic units from the Aegean domain {Jolivet:2008eb}  
729 shows variations in decompression paths similar to synthetic paths computed from  
730 numerical modelling (Figures 10G & 11G), with cooler paths for the latest rocks  
731 exhumed in Crete or in Peloponnese. PT paths shapes are therefore relevantly  
732 reproduced without viscous heating in our models testifying that this process is a  
733 second order phenomenon in detachment system, especially when weak lithologies are  
734 implied over time scales about 10 Myr (Souche et al., 2013) or that shear heating is  
735 counterbalanced by local cooling processes such as fluid advection within detachment  
736 (Famin et al., 2004, Mezri et al., 2015).

737

#### 738 **4.2.3 Temporal evolution of multiple detachment systems**

739 Detachment systems (§ 2.2.2) as well as smaller structures with several detachments  
740 (Betics and Motajica for instance) can be described as a sequence of changes in strain  
741 localization levels. Menderes and Cycladic systems evolve toward the high temperature  
742 end-member in the ternary diagram proposed for natural cases (Figure 5). Nevertheless,  
743 their mechanical evolution differ and they do not follow the same trend in the intra-  
744 crustal strength contrast vs thermal state diagram (Figure 16). In Menderes (Figure  
745 17A), strain is first localized at the base of or within the high-pressure metamorphic  
746 rocks and the Izmir Ankara Ophiolite, along the Simav Detachment, which actually  
747 reactivates an older thrust. Whether the equivalent thrust at the base of the Lycian



748 nappes effectively accommodated extension remains unknown. Even though the  
749 intrusion of the syn-tectonic Egrigoz granite (crystallization at 20-21 Ma and cooling at  
750 20 Ma, Dilek, 2009) is expected to lower the effective strength of the detachment zone,  
751 mylonites along Simav Detachment exhibit similar ages (22.8 Ma, Tekeli, 2001) implying  
752 a short activation duration for the Simav Detachment. Strain is accommodated in a  
753 second time within the lower plate, between 16 and 7 Ma according to syntectonic  
754 granites ages (Figure 17A), basin infills in hanging-walls of detachments (16.7 to 14.9  
755 Ma) and dating of white micas in mylonites ( $7 \pm 1$  Ma Lips, 2001). The Alasehir and  
756 Buyuk detachments (Table 1) constitute a synchronous and antithetic detachment pair,  
757 that relocates strain at the core of the Menderes massif while stretching goes on. This  
758 implies that this domain is eventually weaker than the previously activated Simav  
759 Detachment zone, probably due to crustal thinning, rise of thermal state and intrusion of  
760 granites. Intra-crustal strength contrast is probably rising with thermal state during  
761 stretching and extension mode evolves from perennial asymmetric detachment to a  
762 double-dome symmetric MCC.

763 In Northern Cyclades (Figure 17B) the successive Tinos, Livada and Mykonos  
764 detachment zones (Jolivet et al., 2010) progressively relocate strain higher within the  
765 upper plate while stretching develops from 30 to 9 Ma and thermal state increases with  
766 consecutive intrusions of syn-tectonic granites at 14-15 and 11-13 Ma. Late reactivation  
767 of the Vari detachment on Syros at 10 Ma is limited and the North Cycladic Detachment  
768 System can be considered as a perennial asymmetric detachment system responsible for  
769 most of the exhumation in northern Cyclades. It seems that intra-crustal strength  
770 contrast remained governed by inheritance and that intrusion of granites only promoted  
771 upward migration of the shear zone within the upper plate. For this reason, the North

772 Cycladic Detachment System evolution can be viewed as a rise in thermal gradient  
773 without drastic change in intra-crustal contrast through time.

774 The Betic cordilleras (Figure 17C) clearly show a double detachment system, with the  
775 older Malaguides-Alpujarrides Detachment (activated from 23 to 20 Ma) on top and the  
776 younger Mecina-Filabres Shear Zone (activated from 16 to 7 Ma, Augier et al, 2005).

777 While the first one is highly asymmetric with only top-to-the-North shear sense  
778 indicators developed in the northern mylonitic edge of the Sierra de las Estancias massif  
779 (Lonergan and Platt, 1995), the Mecina Filabres shear zone wraps a multiple dome  
780 structure (Sierra de los Filabres and Sierra Alhamilla) with divergent shear sense  
781 indicators (Augier et al., 2005a). Even if highly tridimensional, this pattern implies that  
782 the older detachment system (MAD, Figure 17C) progressively hardens while lower  
783 detachment activates.

784 In the Carpathian (Figure 17D), the Motajica dome is also the result of a two stage  
785 extension between 25 and 13 Ma (Ustaszewski et al., 2010). Until 18 Ma, the main  
786 structure accommodating strain is the former Tisza-Dacia unit basal thrust reactivated  
787 as a detachment, strain is then accommodated along a symmetric listric faults system  
788 (Figure 17D) that cuts across the Motajica Detachment. The Motajica granite, that  
789 yielded a 27 Ma U-Pb zircon age (Ustaszewski et al., 2010) predates the regional  
790 extension stage so that relocalization of strain within the metamorphic Sava zone cannot  
791 be ascribed to intrusion. Nevertheless, this example is a supplemental illustration of the  
792 recurrent evolution from perennial detachment MCCs toward double-dome MCCs with  
793 increasing stretching of the crust. This conclusion is apparently opposite to the temporal  
794 evolution of extension modes speculated from short term modelling (Buck, 1991) and  
795 exemplified by the Basin and Range: from MCC to wide-rift and then narrow-rift. Ring et

796 al. (2003) also view Menderes as a wide-rift structure overprinting a former MCC  
797 pattern. The main difference between these two evolutions is the heat budget and time  
798 scale. The perennial detachment toward double-dome MCC evolution implies heat input  
799 at the base of the crust higher than heat dissipation due to stretching, due to  
800 lithospheric stretching for instance, while the MCC toward rift evolution implies a  
801 decreasing heat budget due to efficient heat dissipation during MCC extension phase.  
802 While the first evolution seems supported by natural evidences over short time periods  
803 (15 to 20 Myrs in examples developed here, Figure 17), the later could be relevant for  
804 longer time-scale evolutions (over more than 40 Myrs for the Great Basin MCC cluster  
805 (Dickinson, 2002)

806

#### 807 **4.3 Thermal state, inheritance, strain rate and metamorphism as controlling** 808 **parameters for intra-crustal strength contrast**

809 The present study after other numerical modelling of extension modes of the continental  
810 lithosphere (Huet et al., 2010; 2011; Rey et al., 2009; 2011; Tirel et al., 2008; 2006; Wu  
811 et al., 2015) enables to discuss the balance between parameters controlling the intra-  
812 crustal strength contrast within the crust. The present compilation of models shows that  
813 an exaggeration of one order of magnitude of the intra-crustal strength contrast is  
814 sufficient to switch extension mode from rifting to MCC. In a crude approximation, the  
815 impact of the different parameters invoked is to lower the effective strength of the lower  
816 crust. Maximum strength within the crust being controlled by frictional behaviour, it is  
817 unlikely to vary over one order of magnitude for material governed by Byerlee's law.  
818 Using the different lithologies considered as relevant for crustal materials (diorite,  
819 quartzite and wet quartzite), a simple calculation of effective strength for various

820 temperatures and strain rates (Figure 18) shows that for a quartz-diorite at 650°C,  
821 strained at  $10^{-13.5} \text{ s}^{-1}$ , strain rate must decrease of more than 2 orders of magnitude (A in  
822 Figure 18), or temperature must rise of 200°C (B in Figure 18) to lower its effective  
823 strength of 1 order of magnitude. Alternatively, a wet quartzite in the same conditions is  
824 also 10 times less resistant (C in Figure 18). Strain rate and temperature therefore must  
825 change drastically in order to promote high intra-crustal strength contrast. Layering due  
826 to inheritance can therefore be a first order parameter responsible for such contrasts.  
827 These three effects actually can add up: a warm and layered crust stretched at a low  
828 strain rate will definitely develop a MCC pattern.

829 Together with layering, inner complex structures can be inherited from preliminary  
830 thickening stages. Dipping lithological contrast have a strong impact on extensional  
831 behavior of the crust (Huet et al., 2011; Le Pourhiet et al., 2004; Wu et al., 2015). Lateral  
832 strength contrast gradients produced with realistic geometries (Wu et al., 2015) are also  
833 about 1 order of magnitude (Figure 14). Dipping interfaces therefore also constitute a  
834 first order inheritance effect. Transcurrent structures, prone to develop during  
835 thickening

836 Metamorphic reactions responsible for destabilization or crystallization of weak phases  
837 or fluid influx were not taken into account in all the studies considered here, that all  
838 focused on materials with continuous rheology. According to experimental studies  
839 (Jaoul et al., 1984), a quartz aggregate with 0.28 wt% H<sub>2</sub>O has an effective viscosity 100  
840 times lower than a dry aggregate. This severe effect implies that the destabilization or  
841 crystallisation of a limited percentage of hydroxyl-bearing minerals during the  
842 exhumation of the lower plate (Figure 7) can induce high intra-crustal strength contrast.  
843 This effect has also been implemented in numerical modelling with complex feedback

844 between strain, permeability and metamorphic reactions (Mezri et al., 2015). It appears  
845 that models implementing water-controlled phase change develop an asymmetric  
846 detachment MCC evolving into a double-dome MCC, while reference model (TMoho  
847 810°C, stretching rate 0.8 cm/yr) yields a narrow rift extension pattern. Second  
848 invariant of strain rate tensor in the high fluid flux domains is more than one order of  
849 magnitude higher than in the low fluid flux domains, implying a weakening effect of  
850 more than one order of magnitude. Partial melting, observed in most of high  
851 temperature MCCs cores, has also been considered in models as a sharp viscosity drop  
852 within the lower crust (Rey et al., 2009, Schenker et al., 2012). When implemented as  
853 such, solidus envelope at depth constitutes a sharp intra-crustal strength contrast and  
854 promotes the development of double-dome MCCs. Intra-crustal strength contrasts in  
855 these cases are close to 3 orders of magnitude higher than the critical value defined here  
856 (Figure 14, 15).

857

## 858 **Conclusion**

859 The understanding of MCC development and the localization of strain on a perennial  
860 detachment zone was initially based on models with homogeneous crust, and  
861 highlighted the major role of crustal thickness and temperature in the mode of extension  
862 of continental lithosphere. The examination of natural cases in the Mediterranean realm  
863 emphasizes a second type of MCCs exhuming "cold" metamorphic rocks beneath a  
864 detachment system localized along sharp intra-crustal strength contrast inherited from  
865 thickening stages of orogenic crusts. Interlayered HP parametamorphic nappes and  
866 basal contacts of resistant upper units most of the time localize extensional strain. The  
867 comparison of new and published numerical models of extending continental

868 lithosphere confirms the mechanical feasibility of core-complex in continental  
869 lithosphere with low thermal gradients and marked layering. It also shows that intra-  
870 crustal strength contrast is a key parameter for predicting the extension mode of  
871 continental lithosphere. Beyond a critical value of 1000, MCCs develop, whatever the  
872 initial thermal state, thickness or strain rate. For high values of this contrast, double-  
873 dome core complex develop, with strain progressively re-localizing at the center of the  
874 footwall during stretching, while asymmetric dome geometries with one perennial  
875 detachment are produced for intra-crustal strength contrast close to 1000. Natural  
876 examples of Mediterranean detachment systems also show these two MCC types, and  
877 evolution of multiple detachments system show that in most cases inheritance controls  
878 the first extension steps and asymmetric strain pattern, and that further stretching is  
879 accommodated by double dome dynamics, driven by thermal effects mainly.

880 The present study is based on mediterranean that share a common tectonic setting and  
881 possibly some specificities of the post-variscan crust and alpine cover. Nevertheless the  
882 proposed critical parameter for MCC development proposed here, i. e. the intra-crustal  
883 strength contrast and the gradation of extension modes and core-complex types could  
884 be used in comparable settings. The Cordilleran Core Complexes developed during post-  
885 orogenic extension or the numerous comparable studies recently described in China  
886 (Whitney et al., 2013) could also be considered in a broader review.

887

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893 anonymous reviewer for their constructive reviews and careful handling of the present  
894 manuscript.

895

896

897 **Table and Figure captions**

898 **Table 1** : Characteristic features of Mediterranean detachments and associated domes  
899 and basins. Pand T estimates for metamorphic peak conditions (4.2) are given in  
900 supplemental table A1. A: amphibolite, BS: blueschist, E: eclogite; GS: greenschist,  
901 HANF: high angle normal fault, HPA: high pressure amphibolite, HPG: high pressure  
902 granulite, LANF: low angle normal fault, LPA: low pressure amphibolite,

903

904 **Table 2** : Main features of Mediterranean detachments tectonostratigraphies. C for calc-  
905 mylonites, m/c for marine vs continental basins, S and P for syn- and post-orogenic  
906 initiations. c continental, m marine, off off-shore. The relative weighing of ticks between  
907 lines 1,2 and 3, line 4 and line 5 was used to place natural cases on figure 5.

908 AA Alpi Apuane, ABMDS Alasehir-Buyuk Menderes Detachment System, Asp  
909 Aspromonte, ASZ Alakeci Shear Zone, Bal Balagne basal contact, BB Beni Bousera, CaD  
910 Capanne Detachment, CCD Calcare Cavernoso Detachment, CMe Central Menderes, CR  
911 Corinth Rift, CrD Cretan Detachment, Ed Edough, GK Grande Kabylie, Ik Ikaria, Kaz  
912 Kazdağ, KD Kabylia Detachment, KnD Kerdylion Detachment, LL Lubenik Line, MAD  
913 Malguides-Alpujarrides Detachment, MeSZ Messaria Shear Zone, MFSZ Mecina-Filabres  
914 Shear Zone, Mo Motajica, NCDS Northern Cycladic Detachment System, NCy Northern  
915 Cyclades, Ni Niğde, NP Naxos-Paros, NPD Naxos Paros Detachment, SD Simav  
916 Detachment, SD Simav Detachment, Sy Syros, Sif Sifnos, TeSZ Tenda Shear Zone, TKD  
917 Tokachka-Kesebir Detachment, VD Vari Detachment, Ve Veporic, ZFZ Zaouia Fault Zone.

918



919 **Table 3** : Rheological parameters used and varied for the different models. Dislocation  
920 creep law for the mantle is dry dunite (Chopra and Paterson, 1984), and for the crust  
921 quartz-diorite and dry quartzite (Ranalli and Murphy, 1987) and wet quartzite (Kirby  
922 and Kronenberg, 1987).

923 **Table 4** : Simulation results. Each simulation is named after its layering (d: quartz-  
924 diorite, q: quartzite, w: wet quartzite), the initial crust thickness, in km, and initial  
925 bottom temperature in Celsius. Evenly layered crusts have two half-crust layers. All  
926 parameters are computed with initial geometry and a  $1 \text{ cm.yr}^{-1}$  stretching rate for a 210  
927 km long crustal segment, i.e. strain rate =  $2.10^{-15} \text{ s}^{-1}$ . Intra-crustal strength contrast is the  
928 ratio of maximum strength at the brittle-ductile transition over viscous strength at Moho  
929 depth, Integrated lithospheric strength is calculated on the 90 km height of models.  
930 Result types are listed as : R, narrow rifts, WR, wide rifts, PDC perennial detachment  
931 core complexes, DDC double dome core complexes. X refers to geologically meaningless  
932 simulations. Bold lines refer to simulations used as examples in figures 8, 9 and 10.

933 **Figure 1** : Principle sketch of a Metamorphic Core Complex, showing the association of  
934 structures and fault-rock types considered as symptomatic for ductile extensional mode.  
935 Rheological envelope associated to these structures is also shown. BDT : Brittle-ductile  
936 transition.

937 **Figure 2** : Map of Mediterranean core complexes on top of Moho isobaths, homogenized  
938 and smoothed after (Dèzes and Ziegler, 2001; Di Luccio and Pasyanos, 2007; Koulakov  
939 and Sobolev, 2006; Marone et al., 2003; Tiberi et al., 2001) The large Menderes and  
940 Betics domes have been subdivided into units. Shoreline in white.

941 **Figure 3** : Pressure temperature conditions estimated for Mediterranean detachment  
942 footwalls compared to initial Moho temperatures ascribed in numerical models for MCC  
943 formation. Curves 1 : wet solidus of granite (Huang & Wyllie, 1981), 2 and 3 are  
944 maximum and minimum initial thermal gradients tested in the present parametric  
945 study.. AA Alpi Apuane (1 Mass, 2 Autochthon), Alpu Alpujarrides (1 Herradura u.,  
946 2 Sadobreña & Adra u. 3 Escalate u., 4 Lujar Gador u.), And Andros, Asp Aspromonte, BB  
947 Beni Bousera (1 mafic rocks, 2 kinzigites, 3 Filali schists), CC Catena Costiera, Ed  
948 Edough, ERho Eastern Rhodope (1 HP peak, 2 HT,), GK Greater Kabylie (1 Sidi Ali Bou  
949 Nab u., 2 migmatites), Ik Ikaria,, Kaz Kazdag, Me Menderes (1 Cine u. HP , 2 Cine u. HT -  
950 3 Bozdag u., 4 Selimye u.), Mot Motajica, Nax Naxos (1 HP, 2 HT)NF Nevado-Filabrides (1  
951 Bedar Macael u. , 3 Phase D3), Ni Nigde, , Re Rechnitz (1 & 2) SRho Southern Rhodope  
952 (Thasos), Sy Syros (1 high pressure, 2 late overprint) Eclogite peak 2 GS retrogression),  
953 Te Tenda massif, Corsica, Tin Tinos (1 Eclogite peak, 2 BS stage, 3 transition, 4 GS  
954 retrogression), Ve Veporic.

955 P,T values and references are given in supplemental material table A1

956

957 **Figure 4** : Tectonostratigraphic positions of Mediterranean detachments. Colors refer to  
958 metamorphic facies: pale yellow, undistinct low grade, green: GS greenschist, orange: A  
959 amphibolite, brown: G granulite, blue: BS blueschist, purple: E, eclogite. Relevant  
960 lithologies are precised. Extensional detachments are abbreviated as in Table 1.

961 Adria. Adriatic plate, Alpu. Alpujarrides, Aspro. Aspromonte u., APU Amphibolites-  
962 Peridotites u., BMD Buyuk Menderes detachment, CBS Cycladic Blueschists, Eg Egrigoz  
963 granodiorite, FD Fanari Detachment, Ghom. Ghomarides, GT Gavrovo Tripolitza, Gum.

964 Gumusler, HGMU High grade metasedimentary u., HPU High pressure u., IAO Izmir-  
965 Ankara ophiolite, LANF Low Angle Normal Fault, LiD Livada detachment, Malag.  
966 Malaguides, MGU Migmatite-granite u., Mul Mulhacen, MyD Mykonos detachment, N.F.  
967 Nevado-Filabrides, NMM, CMM & SMM North Central and South Menderes massifs,  
968 Pelag. Pelagonian, Phyllites-Q. Phyllites-quartzites, Pi Pindos, Ro Rochovce, S. L. Schistes  
969 Lustrés, Seb. Sebtides, TB Tyros beds, Tell. Tellian u., TiD Tinos Detachment, Tu Turgutlu  
970 granite, UOU, LOU Upper and Lower ophiolite u., ZF Zaroukla fault

971 **Figure 5** : Ternary diagram representing the balance of thermal and inheritance effects  
972 on the different Mediterranean extensional detachments, abbreviated as in Table 1.  
973 Positions in the triangle come from the relative weighting of characteristic features listed  
974 in table 1. Regular case : detachments inherited from syn-orogenic tectonics, bold case :  
975 post-orogenic detachments. HT high temperature footwall, HP high pressure unit  
976 preserved in the footwall, HS high strength hanging wall. White arrows represent time  
977 evolution of detachment systems.

978 **Figure 6** : Synoptic comparative time evolution or relative position of syn-detachment  
979 collapse basins preserved in the Mediterranean realm according to their sedimentary  
980 infill. Arrow senses represent time evolution : most of basins evolve toward marine  
981 (downward arrows), only NaPaD basin is evolving from marine to continental with time  
982 (upward bold arrow). Color represent paleobathymetry for marine basins (light grey:  
983 shallow, darker grey: deep) Refers to line 2.1 in table 1 and references therein.  
984 Abbreviations as in table 1.

985 **Figure 7** : Thermochemical modelling of mode evolution of three protoliths  
986 representative for pelites, granites and basalts as detailed in table I B95, Boggs, 1995,  
987 LM76 : Le Maitre, 1976. A, B, C and E,F,G : modal evolutions in volume fractions along

988 paths 1 and 2 as shown on J. Micas are phengite and paragonite together. More chaotic  
989 curves in C and G are due to poorly constrained complex local equilibria along main  
990 reactions in mafic rocks. D and H : Phyllosilicates volume fraction evolution along the  
991 same paths. Shaded areas represent brittle quartz domain.

992 **Figure 8** : Set-up of numerical experiments initial conditions for a 45 km thick double  
993 layered crust (d/q\_45\_1200 cases in table 4) See Huet et al., 2011 for details. Bottom  
994 temperatures are varied between 1000, 1200 and 1400°C. Resulting initial Moho  
995 temperatures are listed in table 3. Hatched column on the left represent a 20 km wide  
996 area with no noise on cohesion and a higher cohesion (30 MPa instead of 20 MPa).

997 **Figure 9** : Compared output of numerical experiments for simulation d/q\_30\_1400,  
998 representative of a rift-type extension mode (A & C) after 6 Myr stretching and  
999 simulation q/q\_45\_1200 representative for the wide rift extension mode after 12 Myr  
1000 stretching. White curves show foliation trajectories from finite strain computed as in  
1001 Huet et al., 2011. C & D represent instantaneous strain rates. Initial strength profiles  
1002 computed for  $2 \cdot 10^{-15} \text{ s}^{-1}$  and parameters detailed in table 2 are depicted in E and F.  
1003 Values are precised in table 3.

1004 **Figure 10** : Simulation d/q\_45\_1200 output representative for perennial detachment  
1005 core complex extension mode. A & B : geometry after 4 and 14 Myr stretching, with  
1006 position of markers used for PT and exhumation calculations. White curves show  
1007 foliation trajectories from finite strain computed as in Huet et al., 2011. C & D :  
1008 instantaneous strain rates profiles, E Initial strength profile computed for  $2 \cdot 10^{-15} \text{ s}^{-1}$  and  
1009 parameters detailed in table 2F & G: exhumation history and PT paths for markers 1, 2  
1010 and 3.

1011 **Figure 11** : Simulation d/w\_45\_1200 output representative for double dome core  
1012 complex extension mode. A & B: geometry after 4 and 14.1 Myr stretching, with position  
1013 of markers used for PT and exhumation calculations. White curves show foliation  
1014 trajectories from finite strain computed as in Huet et al., 2011. C & D : instantaneous  
1015 strain rates profiles, E Initial strength profile computed for  $2 \cdot 10^{-15} \text{ s}^{-1}$  and parameters  
1016 detailed in table 2F & D: exhumation history and PT paths for markers 1, 2 and 3.

1017 **Figure 12** : Resulting geometries of the simulations run in this study. Series are referred  
1018 to as in table 3. A, B & C represent homogeneous crust simulations, D and E evenly  
1019 layered crust simulations and F unevenly layered simulations. Line 45 km on D  
1020 represents the same simulations as line 1/2 on F.

1021 **Figure 13** : Intra-crustal strength contrast vs integrated lithosphere strength logarithm  
1022 plot for all 48 meaningful simulations listed in table 3. Effect of different tested  
1023 parameters is highlighted by series tied by dash lines. Triangles: series d/q\_X\_1000,  
1024 squares: series d/d\_60\_X, diamonds: d/X\_45\_1200, crosses: d/q\_45\_1200. 1/3, 1/2, 2/3  
1025 refer to the relative size of the upper crust in unevenly layered crust simulations.

1026 **Figure 14** : Intra-crustal strength contrast vs integrated lithosphere strength log-log  
1027 plot for 46 published numerical simulations compared to present study. TS This Study,  
1028 H11 (Huet et al., 2011), all 9 simulations are reported, T08 (Tirel et al., 2008) all 32  
1029 simulations reported, R09 (Rey et al., 2009) 2 simulations reported, with no melt effect  
1030 at slow ( $2 \cdot 10^{-16} \text{ s}^{-1}$ ) and fast ( $2 \cdot 10^{-15} \text{ s}^{-1}$ ) stretching rate. S12 (Schenker et al., 2012) all 12  
1031 models reported. W15 (Wu et al., 2015) 3 models reported, with wedge-shape layered  
1032 crust and no prescribed weak interface. Calculations were made at each initial model  
1033 end (thick and thin ends). Bold squares : stretching rate effect, as explored by Tirel et al.

1034 (2008) on the series with initial crustal thickness 60 km and initial Moho temperature  
1035 830°C (mantle heat flux 30mW.m<sup>-2</sup>)

1036 **Figure 15** : Intra-crustal strength contrast vs initial Moho temperature logarithm plot  
1037 for 46 published numerical simulations compared to present study. Legend as in figure  
1038 14. Triangle inset refers to ternary end-member MCC types analogue to ternary diagram  
1039 used for Mediterranean examples (Figure 5). Symbols at triangle edges refer to where  
1040 different studies plot in this diagram.

1041 **Figure 16** : Synthetic diagram showing preponderance of intra-crustal strength  
1042 contrast over thermal state of the crust on extension mode in mediterranean natural  
1043 cases on the basis of thermal evolutions recorded.

1044 **Figure 17** : Examples of time evolution of multiple detachment systems in the  
1045 Mediterranean. Detachment named as in Table 1. Moho depth as in Figure 2. A  
1046 Menderes, IAO : Izmir Ankara Ophiolite, modified after VanHinsbergen, 2010 and Ring,  
1047 2003, B- North Cycladic Detachment System, after Jolivet et al., (2010), C Betic  
1048 Cordilleras after Augier et al. (2005) and Lonergan and Platt (1995), D Motajica after  
1049 Ustazjewski et al. (2010). g. for granite

1050 **Figure 18** : Strength contours in MPa (presented with a logarithmic scale) for different  
1051 crustal lithologies used in the present study. A, B and C are responsible for the same  
1052 strength drop of 1 order of magnitude. A: strain rate drop, B: temperature increase, C:  
1053 lithological change from quartz-diorite to wet quartzite.

1054

1055

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<b>1. Locality</b>		<b>A- Betics, Spain</b>		<b>B- Beni Bousera (B.B.), Morocco</b>
	2.1 Syntectonic basin		continental to marine (de Galdeano and Vera, 1992)	
	2.2 Basal lithologies	<b>Malaguides</b> : palaeozoic sandstones and slates	<b>Alpujarrides</b> : phyllites, quartzites and graphitic micaschists	<b>Ghomarides</b> : paleozoic metasediments
<b>2. Upper plate</b>	2.3 Peak metamorphic conditions	lower anchizone (variscan)	HPA in Herradura to GS in Lujar Gador (Azañón and Crespo-Blanc, 2000; Jolivet et al., 2010b)	GS ?
	2.4 Brittle deformation		2 sets of LANF (Platzman and Platt, 2004; Augier et al., 2005)	
	3.1 Abbreviation/name	<b>MAD/Malaguides-Alpujarrides Detachment</b> (Platzman and Platt, 2004)	<b>MFSZ/Mecina-Filabres Shear Zone</b> (Platt et al., 1984)	<b>ZaFZ/Zaouia Fault Zone</b>
<b>3. Detachment Zone</b>	3.2 Associated lithologies	Calc-mylonites and fault gouge (Lonergan and Platt, 1995)	mylonites and gouges	« brittle-ductile fault » (Michard et al., 2006)
	3.3 Relationships to heritage	reactivated thrust	reactivated thrust	oblique to thrusts
	4.1 Top lithologies	<b>Alpujarrides</b> : phyllites, quartzites and graphitic micaschists	<b>Nevado-Filabrides</b> : permo-triassic graphitic schists (Bedar Macael)	<b>Sebtides</b> : permo-triassic metasediments (Filali schists & Federico)
<b>4. Lower plate</b>	4.2 Peak metamorphic conditions*	HPA in Herradura (Alpu1) to GS in Lujar Gador (Alpu4) (Azañón and Crespo-Blanc, 2000)	E in Bedar Macael (NF1) to GS in Ragua (NF3, Augier et al., 2005)	HPA in Federico (BB3) HPG in B.B. kinzigites (BB2)
	4.3 Brittle deformation			
<b>5. Migmatites</b>		at the base of Herradura	no HT overprint	
	6.1 Type	mafic dikes in upper and lower plates		leucogranite dikes in B.B. lower u.
<b>6. Intrusives</b>	6.2 Relation to ductile deformation	late ? magmatism		undeformed
<b>7. Ophiolitic units</b>	7.1 Type			peridotites
	7.2 Position			in the dome core
<b>8. Timing in orogeny</b>		syn-orogenic initiation (Jolivet et al., 2008)	mainly post-orogenic (Jolivet et al., 2008)	possibly syn-orogenic (Michard et al., 2006)
<b>9. Geodynamic boundary conditions</b>		change of slab retreat direction from southward to westward		Alboran slab retreat

Table 1a

<b>1.</b>	<b>C- Grande Kabylie, Algeria</b>	<b>D- Edough, Algeria</b>	<b>E- Aspromonte, S. Italy</b>	<b>F- Catena Costiera, S. Italy</b>
2.1	Kabylian oligo-miocene marine conglomerates (OMK)	none	continental to marine Stilo-Capo d'Orlando Formation (SCOF) (Bonardi et al., 1980)	Miocene to Q. marine b. (e. g. Amantea b. (Mattei et al., 2002)
<b>2.</b>	pre-permian phyllites	Tellian and Green Schist facies Unit GFU (Caby et al., 2001)	<b>Stilo u.</b> : para- and orthogneiss	Upper Ophiolitic U. (UOU) : ophiolitic melange & Verbicaro carbonates (Bonardi et al., 2001)
2.3	variscan GS	GS	variscan GS to A, no Alpine overprint (Heymes et al., 2008)	GS
2.4	normal faults in Central and Eastern Grande Kabylie			tilted blocks at the base of basin
3.1	<b>KaD/Kabylian Detachment</b> (Saadallah and Caby, 1996)	Ed/~KaD, major metamorphic gap	Asp/	CatCo/
<b>3.</b>	mylonites and younger cataclastes	GS mylonites, fibrolites bearing ultra-mylonites at base of APU	mylonitic gneiss	mylonites and cataclastes (Rossetti et al., 2004)
3.3		reactivated Tellian thrust	reactivated thrust	
4.1	<b>Sidi Ali Boud Nab</b> massif: Micaschists and orthogneiss	high grade metasedimentary U. (HGMU) Amphibolite-Peridotite U. (APU) Migmatite-Granite U. (MGU)	<b>Aspromonte u.</b> : para and orthogneiss	Continental metaclastics and marbles Lower Ophiolitic U. (LOU)
4.2	A (GK1 & GK2)	HPG	Alpine BS over variscan LPA	BS (Rossetti et al., 2004)
4.3				
<b>5.</b>	anatectic metapelites	syn-tectonic anatexis in MGU	none	none
<b>6.</b>	6.1 aplo-pegmatites and leucogranites	leucogranites	none	none
6.2	syn-kinematic aplites / late granites	syn-kinematic in MGU		
<b>7.</b>	7.1	peridotites at the base of APU (Bosch et al., 2014)	none	ophiolitic melanges, pillow-lavas and picrites in Verbicaro
7.2		top lower plate and lenses in core		in lower and upper plates
<b>8.</b>	post-orogenic	late post-orogenic	syn-to immediate post-o. (Heymes et al., 2010)	post-orogenic (Jolivet et al., 2008)
<b>9.</b>	Alboran slab retreat	Alboran slab retreat or later slab dynamics (Bruguier, 2009)	stretching oblique to inferred slab retreat direction	E-W stretching parallel to local slab retreat in miocene

Table 1b

<b>1.</b>	<b>G- Alpi Apuane, N. Italy</b>	<b>H- Elba, Thyrrhenian sea</b>	<b>I- Corsica, W. Mediterranean</b>	
2.1	marine « Breccie delle Versilia » (Carmignani and Kligfield, 1990)	-	-	St Florent b. continental to marine limestones (Ferrandini et al., 1998)
2.2	<b>Tuscan nappes</b> : carbonates with flysch on top	Upper Toscan ophiolite and flysch u.	<b>Balagne-Nebbio unit</b> : ophiolitic unit	<b>Schistes Lustrés (SL)</b> : calcschists, ophiolitic and continental basement slices
2.3	no alpine metamorphism	weak thermometamorphism	non metamorphic	BS to E
2.4	H & LANF rooting on detachment	<b>Zuccale Fault</b> : LANF with cataclasites & calcmylonites crosscuts CaDZ (Collettini and Holdsworth, 2004)	late normal faults (D5, Marroni and Pandolfi, 2003)	
3.1	<b>CalCa/Calcere Cavernoso</b> (Carmignani and Kligfield, 1990)	<b>CaDZ/Capanne Detachment Zone</b> (Daniel and Jolivet, 1995)	Bal/	<b>TeSZ/Tenda Upper Shear Zone</b> (Jolivet et al., 1990; Gueydan, 2003; Fournier et al., 1991)
3.2	mylonites & cataclasites from anhydrite and dolostone	mylonites within thermometamorphic aureole		mylonites
3.3	reactivated thrust	oblique to thrusts	reactivated basal thrust (D4, Marroni and Pandolfi, 2003)	reactivated basal thrust
4.1	<b>Massa &amp; « Autochthonous » u.</b> : paleozoic phyllites and metasandstones	<b>Lower Tuscan u.</b> : Calamite paleozoic schists	<b>Schistes Lustrés (SL)</b> : calcschists, ophiolitic and continental basement slices	Tenda massif : palaeozoic granitoid
4.2	BS (AA1 and AA2)	GS and thermometamorphism	BS to E	BS (Te)
4.3				
5.	none	none	none	none
6.1	volcanism 7 Ma younger	granodiorite (Monte Capanne)	Sisco lamproïtes in SL	
6.2	(Carmignani et al., 2001)		18 Ma younger than extension (Beccaluva et al., 2004)	
7.1	Liguride ophiolites		Nebbio and SL ophiolites	
7.2	top of upper plate, limited extensional reactivation		uppermost (Balagne)	
8.	post-orogenic		post-orogenic	
9.	Thyrrhenian slab retreat		Thyrrhenian slab retreat	

Table 1c



1.	<b>J- Motajica</b> window, Bosnia and Hercegovina	<b>K- Rechnitz window</b> area, Austria/Hungary	<b>L- Central Western Carpathians</b> , Slovakia	<b>M- Southern Rhodope</b> , Greece
2.1	Ottngangian syn-rifts sediments : alluvial to marine } (Pavelić, 2001)	Fürstenfeld (W) and Danube (E) basins : transitional Karpatian deposits (Piller et al., 2007)		
2.	<b>Sava Zone</b> metasediments and ophiolites at the base of Tisza-Dacia u. (Ustaszewski et al., 2010)	<b>Wechsel u.</b> (base of Auto-Alpine) gneiss and Palaeozoic carbonates (Cao et al., 2013)	<b>Gemic u.</b> : palaeozoic basement rocks	<b>Vertiskos Gneiss Complex</b> : gneiss + <b>Therma and Volvi gabbros</b>
2.3	A to GS	GS	low grade	(U)HP (Liati, 2005)
2.4	late brittle normal faults	HANF		upper Miocene faulting
3.1	<b>MoD/Motajica Detachment</b> (Ustaszewski et al., 2010)	ReW/	<b>LuL/Lubenik Line</b> (Janák et al., 2001)	<b>KerD/Kerdylion Detachment</b> (Brun and Sokoutis, 2007)
3.	Upper GS foliation in paragneiss	Serpentinite and quartz-micas mylonites (Cao et al., 2013)	mylonites from cover u. and topmost intrusive granitoid	mylonites and ultra-mylonites
3.3	Partly reactivated thrust	possibly reactivated basal AA thrust	reactivated thrust	
4.1	Accretionary wedge complex and Motajica granite	<b>Rechnitz ophiolite u.</b> Penninic greenschist and serpentinites.	<b>Veporic u.</b> : pre-Alpine basement and phyllite cover	<b>Southern Rhodope Core Complex</b> : marbles and orthogneiss
4.	A 5mot) and local contact metamorphism	BS (Re1) and subsequent GS (Re2)	A (Ve) and contact metamorphism	A (Barrovian gradient SRho)
4.3		Chlorite breccias in borehole		upper Miocene faulting
5.	none	none	hercynian remnants	in the Kerdylion massif
6.1	S- to I-type Motajica granite (Pamić et al., 2012)	none	Rochovce granite	Vrondou and Symvolon granodiorites
6.2	crystallization 5 Myrs prior to extension onset		syn-tectonic	syn-kinematic
7.1	Western Vardar ophiolite complex	Serpentinite, metagabbro, ophicarbonates	none	Volvi and Therma metabasites : gabbros and basalts
7.2	Upper level of upper plate	within lower plate		base of the upper plate
8.	post-orogenic Pannonian extension	post-orogenic Pannonian extension ?	Cretaceous syn-orogenic	post-orogenic
9.	Carpathian subduction roll-back (Royden, 1993)	Eastward lateral extrusion (Ratschbacher et al., 1991) and Carpathian roll-back	Growth of collision wedge after Meliata suture	

TableId

1.	N- Eastern Rhodope, Greece	O- Northern Cyclades, Greece (Andros-Tinos-Mikonos-Ikaria)	P- Syros, Greece	Q- Central Cyclades, Greece (Naxos Paros)
2.1	<b>Krumovgrad</b> grp : colluvial to marine (Bonev et al., 2006)	marine molasses then conglomerates on Mykonos (Sanchez-Gomez et al., 2002)		from marine sandstones and pelites toward fluvial deposits (Kuhlemann et al., 2004; Bargnesi et al., 2013)
2.	<b>Kimi Complex</b> : high grade rocks + meta-ophiolitic lenses.	<b>Upper Cycladic Nappe</b> : ophiolitic complex (UCN)	<b>Vari unit and "Upper Unit"</b> metabasites, gneiss and serpentinites (Soukis and Stockli, 2013)	ophiolite nappe and neritic limestones
2.3	(U?)HP 73 Ma (Krohe and Mposkos, 2002)	GS	HT (Jolivet and Brun, 2008)	unmetamorphosed
2.4	HANF offsetting detachment	HANF branching on detachment	late brittle LANF	synthetic normal faults
3.	<b>ToKeD/Tokachka and Kechros detachments</b>	<b>NCDS/North Cycladic Detachment System</b> (Jolivet et al., 2010a)	<b>VaD/Vari Detachment</b>	<b>NaPaD/Naxos Paros Detachment</b> (Gautier et al., 1993)
3.1				
3.2	mylonites	mylonites, cataclasites and breccias	mylonites and late breccias	mylonites and ultra-mylonites
3.3		partly reactivating UCN thrust		reactivated thrust
4.	<b>Kesebir-Kardamos and Kechros complexes</b> : migmatitic gneiss and orthogneiss (Bonev et al., 2006)	<b>Cycladic Blueschist Unit</b> : metapelites and marbles on granitic basement	Cycladic Blueschist Unit : metapelites, marbles metaflysch (Bonneau, 1984)	Pre-Alpine Gneiss basement with marbles and schists.
4.1				
4.2	lower E (ERho1, Mposkos, 1998)	E-BS (Tin1 to 3) then A (Mykonos and Ikaria; Ik)	E (Syr1)-BS (Syr2) (Trotet et al., 2001)	E then A (Nax 1&2, Avigad, 1998)
4.3	HANF offsetting detachment			
5.	in Kardamos complex	on Dilos, Mykonos and Ikaria	none	pervasive melting (Vanderhaeghe, 2004)
6.	Papikion granitoid	I-type granites on Tinos, Mykonos and Ikaria (JOLIVET)	none	granodiorite (Naxos) and leucogranites (Paros)
6.1				
6.2	presumably syn-tectonic	syn- to late-tectonic		syn-extensional leucogranite
7.	Metaophiolite lenses	serpentinites, gabbros and basalts	metabasites of the Upper Unit	gabbros and ultramafics (Jansen, 1977)
7.1				
7.2	at the base of upper plate (Burg et al., 1996)	at the base of the upper plate	below early then above late detachment (Soukis and Stockli, 2013)	at the base of the upper plate
8.	syn-orogenic	post-orogenic	mainly syn-orogenic	post-orogenic
9.	Vardar ocean closure	Aegean back-arc extension		Aegean back-arc extension

Table e

1.	<b>R- SouthWestern Cyclades, Greece</b> (Kea-Kythnos-Serifos)	<b>S- Kazdag, Turkey</b>	<b>T- Menderes, Turkey</b>	
2.1	possibly on Milos	<b>Küçükkuyu Fm</b> : fluvio-lacustrine to turbidites (Bonev and Beccaletto, 2007)	Late deposits only	lacustrine to shallow marine supra-detachment basins (Sümer et al., 2013)
2.	<b>Pelagonian unit</b> : serpentinite and schists	<b>Cetmi melange</b> : ophiolitic melange (Beccaletto and Steiner, 2005)	Ismir-Ankara Ophiolite, HP-LT nappes at the base	Northern and Southern Menderes
2.3	A-GS	E blocks in the melange	E-BS in Afyon, Tavsanli and Oren nappes (Plunder et al., 2013)	A
2.4	HANF	HANF crosscutting detachment	HANF bounding Simav graben	Half-grabens on both sides
3.1	<b>WCDS/Western Cycladic Detachment System</b> (Grasemann et al., 2011)	<b>AlSeD/Alakeçi SZ (N) and Selale Detachment (S)</b>	<b>SiD/Simav Detachment</b> (Isik and Tekeli, 2001)	<b>AlBuD/Alasehir &amp; Buyuk Detachments</b>
3.2	(ultra)mylonites from marbles and schists	mylonites, metaserpentinites and breccias	mylonites	rmylonites only along Alasehir (Isık et al., 2003) abundant cataclasites
3.3	reactivated thrusts ?	thrust ?	reactivated thrusts	cut through Menderes units
4.1	<b>Cycladic Blueschist Unit</b>	High grade basement	<b>Northern Menderes</b> : Selimiye metasedimentary unit on top	Central Menderes
4.2	BS then GS	A (Kaz, Okay & Satir, 2000)	A (Me4)	A
4.3	HANF			
5.	none	at the top of the dome	inherited in the core (Schuiling, 1962)	?
6.1	I-type granodiorite (Serifos and Lavrion)	Evciler calc-alkaline granite	Early northern MEG	Later central MEG
6.2	syn- to late tectonic	intrudes ASZ	syntectonic to SiD (Isik et al., 2004)	syn-extensional (Dilek et al., 2009)
7.1	serpentinites	Cetmi, Karakaya, and Denizgören	Ismir-Ankara Ophiolite	none
7.2	within the upper plate	at the base of the upper plate	Within the upper plate	
8.	post-orogenic	post-orogenic	post-orogenic	later post-orogenic
9.	Aegean back-arc extension	Anatolian back-arc extension	Eastern termination of NCDS (JOLIVET)	Intra-Menderes stretching (van Hinsbergen, 2010)

Table f

	<b>U- Corinth Rift, Greece</b>	<b>V- Crete, Greece</b>	<b>W- Nigde, Turkey</b>
<b>1.</b>	fluvio-lacustrine toward marine (Rohais et al., 2007)	possibly lacustrine breccias at the base of Neogene basins (Jolivet et al., 1996)	shallow marine deposits in the late Ulukisla basin (Gautier et al., 2008)
<b>2.</b>	<b>Gavrovo-Tripolitza (GT)</b> carbonates with <b>Tyros beds (TB)</b> shales at the base	<b>Gavrovo-Tripolitza</b> carbonates	Massive ophiolite, mostly gabbros (Goncuoglu et al., 1991)
	GS (Jolivet et al., 2010b)	unmetamorphosed	GS at the base
	Zaroukla cataclastic devollement between GT and TB (Sorel, 2000)	HANF cutting across DZ	
<b>3.</b>	<b>CrD1/Cretan Detachment</b> (Jolivet et al., 2009)	CrD2/ (Kilias et al., 1994)	NiD
	high strain phyllites	mylonites from the upper plate	mylonites to ultracataclasites
	reactivated thrust	reactivated thrust	recativated obduction thrust
<b>4.</b>	<b>Phyllites-Quartzites</b>	<b>Phyllites-Quartzites</b> nappe on top of Plattenkalk	<b>Gumusler Formation</b> : migmatites
	BS	BS	A (Nig, Whitney and Dilek, 1997)
	Rift faults cutting across detachment		
<b>5.</b>	none	none	most of the lower plate
<b>6.</b>	none	none	Uçkapili leucogranite
			mainly post-tectonic (Gautier et al., 2002)
<b>7.</b>	none	ophiolite Asterousia nappe on top of upper plate	massive ophiolite in the upper plate, dilascerated slivers in the lower plate
<b>8.</b>	syn- to post-orogenic	syn-orogenic (Jolivet et al., 1996)	syn-orogenic ?
<b>9.</b>	early Aegean extension	Aegean back-arc extension	Central Anatolian edge extension

Table1g



	<b>Crust</b>			<b>Mantle</b>
<b>density at 25°C (kg.m<sup>-3</sup>)</b>		2700		3300
<b>thermal conductivity (W.m<sup>-1</sup>.K<sup>-1</sup>)</b>		2.5		3.3
<b>Thermal expansion coefficient (K<sup>-1</sup>)</b>			3.0 x 10 <sup>-5</sup>	
<b>Specific heat (J.kg<sup>-1</sup>.K<sup>-1</sup>)</b>			10 <sup>3</sup>	
<b>Shear modulus (GPa)</b>			30	
<b>Poisson's ratio</b>			0.25	
<b>Mohr-Coulomb criterion</b>			20	
<b>Cohesion (MPa)</b>			20	
<b>Friction angle (°)</b>			30	
<b>A (MPa<sup>-n</sup>.s<sup>-1</sup>)</b>	1.26 10 <sup>-3</sup>	6.7 10 <sup>-7</sup>	2.51 10 <sup>-4</sup>	2.5 x 10 <sup>4</sup>
<b>n</b>	2.4	2.4	2.3	3.5
<b>E (kJ.mol<sup>-1</sup>)</b>	219	156	154	532
<b>Lithology</b>	quartz-diorite	quartzite	wet quartzite	dunite

table 3

<b>Model</b>	<b>Ductile-brittle transition depth</b>	<b>Initial Moho temperature</b>	<b>Intra-crustal Strength Contrast</b>	<b>Integrated Lithospheric Strength</b>	<b>Extension Mode</b>
units	[km]	[°C]		[MPa x km]	
<b>Homogeneous crust :</b>					
upper/lower crust _thickness [km]_Tbottom [° C]    d: quartz-diorite, q: quartzite, w: wet quartzite					
d/d_30_1000	26.4	398	4	33739	R
d/d_30_1200	22.3	477	17	20860	R
d/d_30_1400	19.4	557	61	12846	R
d/d_45_1000	27.5	569	103	23239	X
d/d_45_1200	23.3	683	413	9688	WR
d/d_45_1400	20.3	797	1232	5437	WR
d/d_60_1000	28.7	725	822	11564	WR
d/d_60_1200	24.3	870	2832	7244	DDC
d/d_60_1400	21.1	1015	7304	5426	DDC
q/q_30_1000	18.4	398	31	30412	R
q/q_30_1200	15.7	477	93	18245	R
q/q_30_1400	13.7	557	223	10834	R
q/q_45_1000	19.2	569	350	19165	WR
<b>q/q_45_1200</b>	<b>16.4</b>	<b>683</b>	<b>914</b>	<b>6761</b>	<b>WR</b>
q/q_45_1400	14.3	797	1924	3227	WR
q/q_60_1000	20.0	725	1559	7119	DDC
q/q_60_1200	17.0	870	3618	4062	DDC
q/q_60_1400	14.9	1015	6918	3026	DDC
w/w_30_1000	15.3	398	135	29119	R
w/w_30_1200	13.0	477	414	17274	WR
w/w_30_1400	11.4	557	1029	10079	WR
w/w_45_1000	16.0	569	1616	17720	WR
w/w_45_1200	13.6	683	4348	5697	DDC
w/w_45_1400	11.9	797	9443	2405	DDC
w/w_60_1000	16.6	725	7478	5554	DDC
w/w_60_1200	14.1	870	17909	2912	DDC
w/w_60_1400	12.3	1015	15913	2138	DDC
<b>Even layered crust</b>					
d/q_30_1000	18.4	398	31	30412	R
d/q_30_1200	15.7	477	93	18245	R
<b>d/q_30_1400</b>	<b>15.1</b>	<b>557</b>	<b>244</b>	<b>10950</b>	<b>R</b>
d/q_45_1000	22.6	569	409	19757	WR
<b>d/q_45_1200</b>	<b>22.5</b>	<b>683</b>	<b>1229</b>	<b>8313</b>	<b>PDC</b>
d/q_45_1400	20.3	797	2664	5123	PDC
d/q_60_1000	28.7	725	2195	10504	DDC
d/q_60_1200	24.3	870	5060	7066	DDC
d/q_60_1400	21.1	1015	9564	5385	DDC
d/w_30_1000	15.3	398	135	29119	X
d/w_30_1200	15.1	477	475	17506	R

d/w_30_1400	15.1	557	1329	10688	R
d/w_45_1000	22.6	569	2232	19419	X
<b>d/w_45_1200</b>	<b>22.5</b>	<b>683</b>	<b>6943</b>	<b>8217</b>	<b>DDC</b>
d/w_45_1400	20.3	797	15448	5089	DDC
d/w_60_1000	28.7	725	12530	10419	DDC
d/w_60_1200	24.3	870	29764	7041	DDC
d/w_60_1400	21.1	1015	26204	5375	DDC

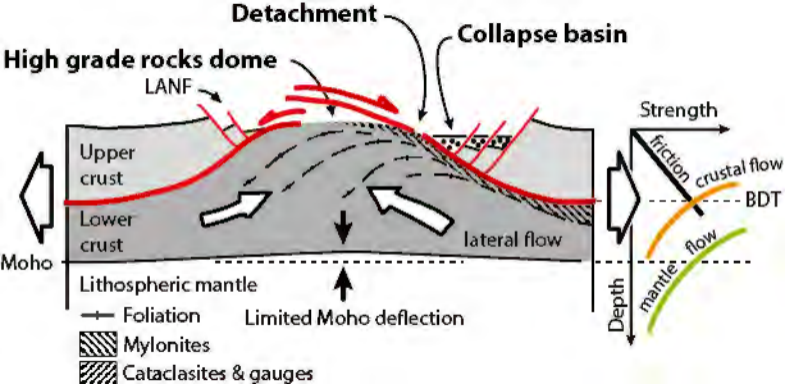
**Uneven layered crust :**

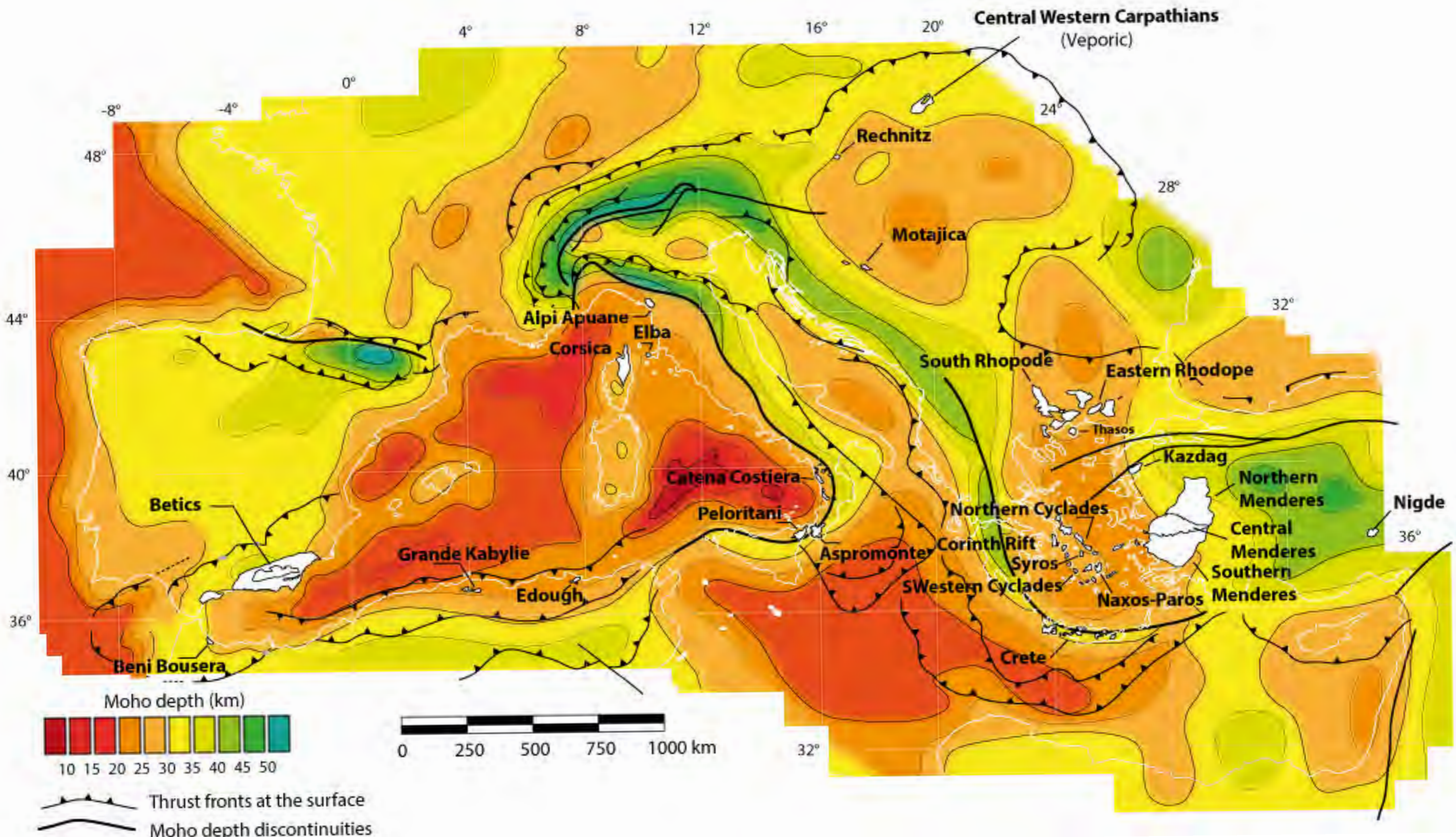
upper crust/lower crust\_upper thickness/total thickness [km]\_Tbottom [°C]

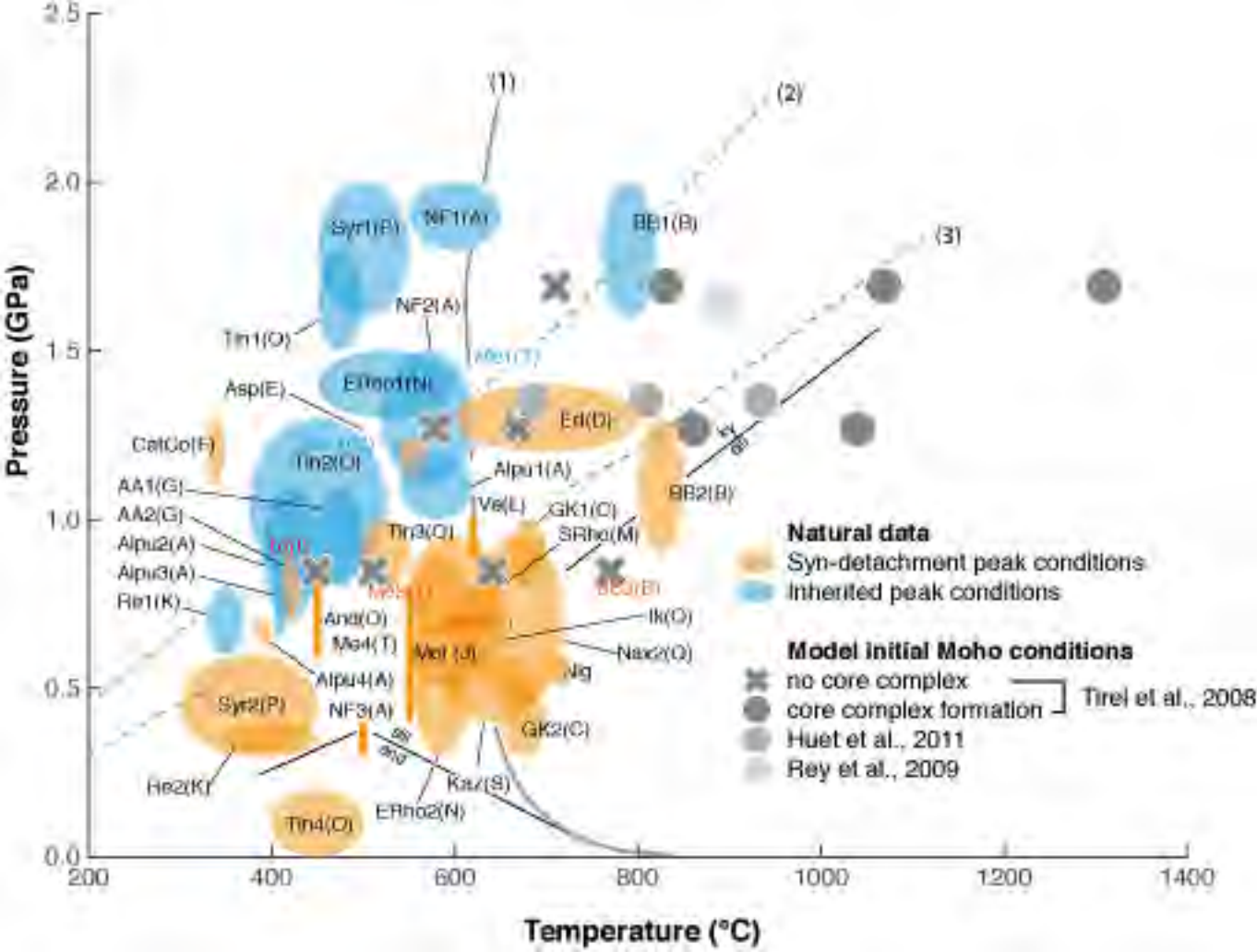
d/q_15/45_1000	19.2	569	350	19165	WR
d/q_15/45_1200	16.4	683	914	6761	WR
d/q_15/45_1400	15.1	797	2023	3268	DDC
d/q_30/45_1000	27.5	569	492	22579	R
d/q_30/45_1200	23.3	683	1270	9576	PDC
d/q_30/45_1400	20.3	797	2664	5411	PDC

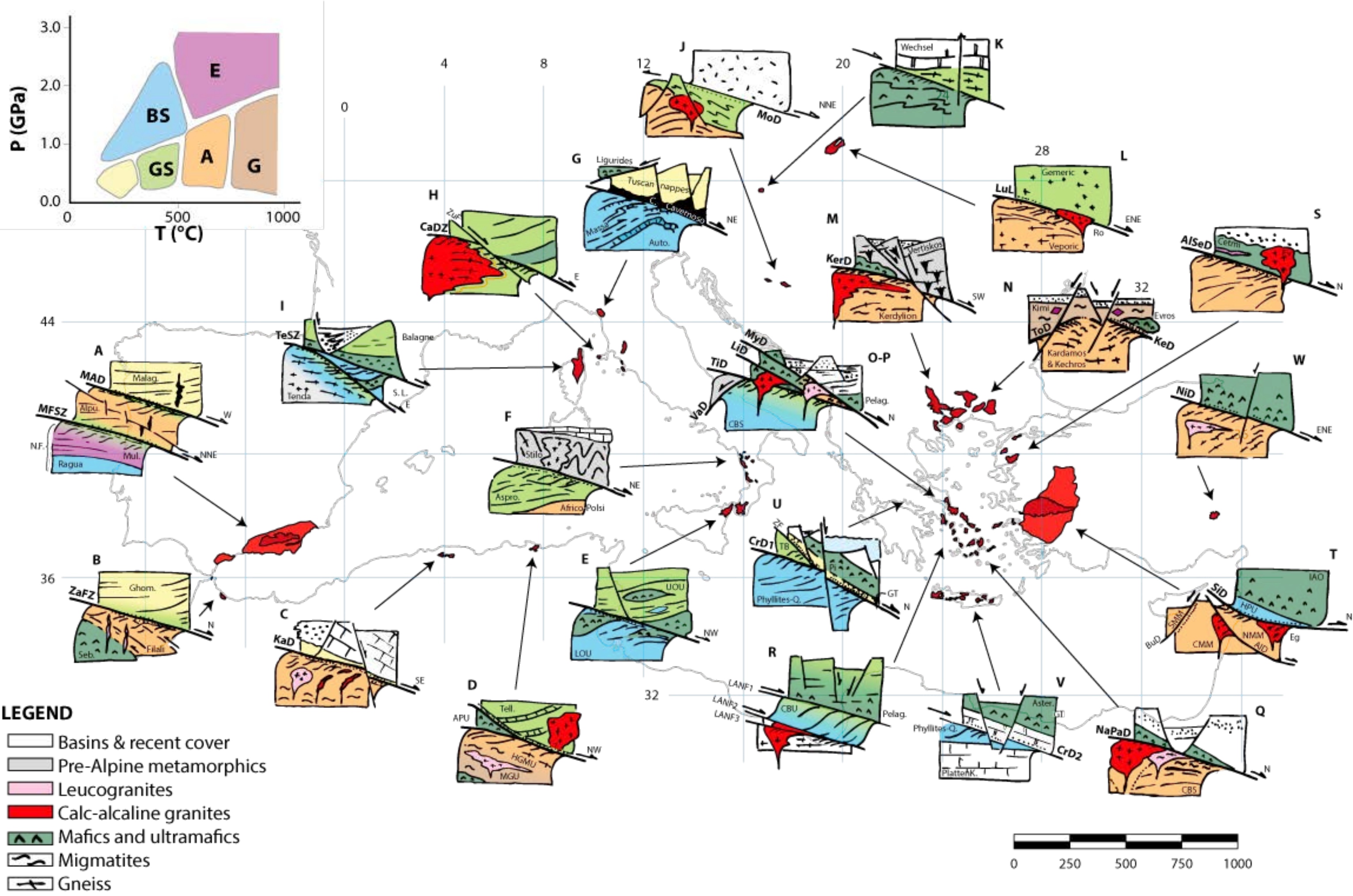
table 4

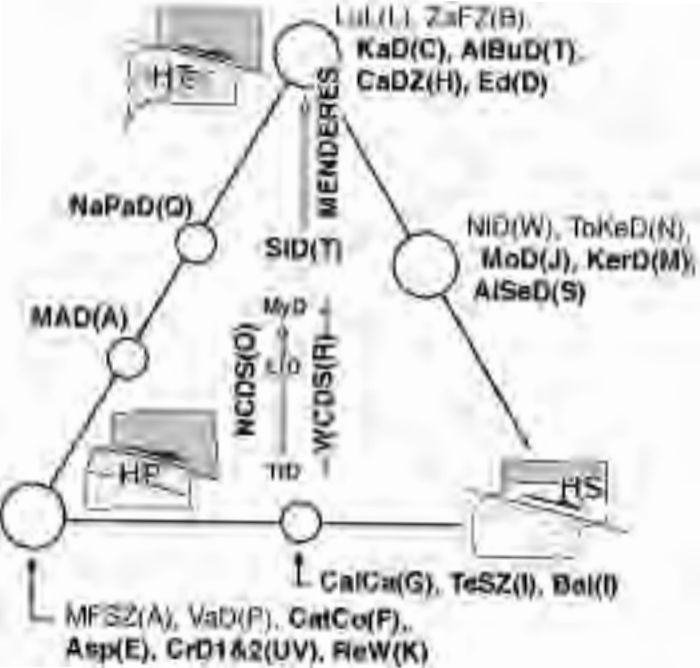


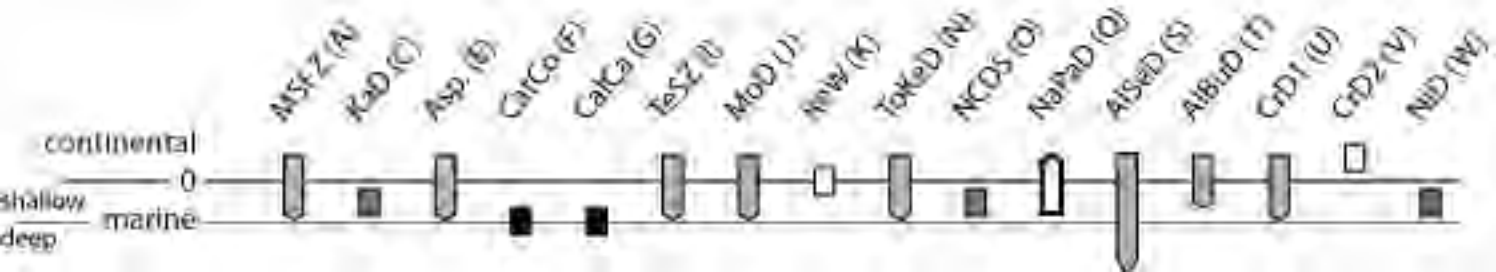




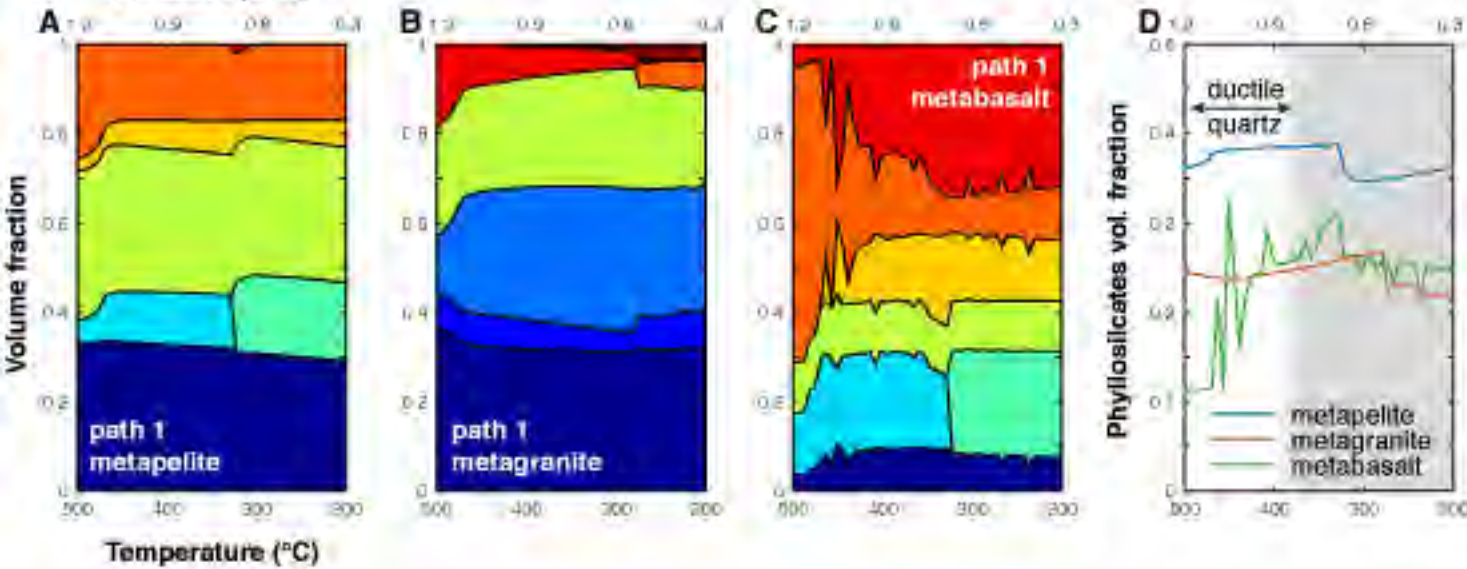




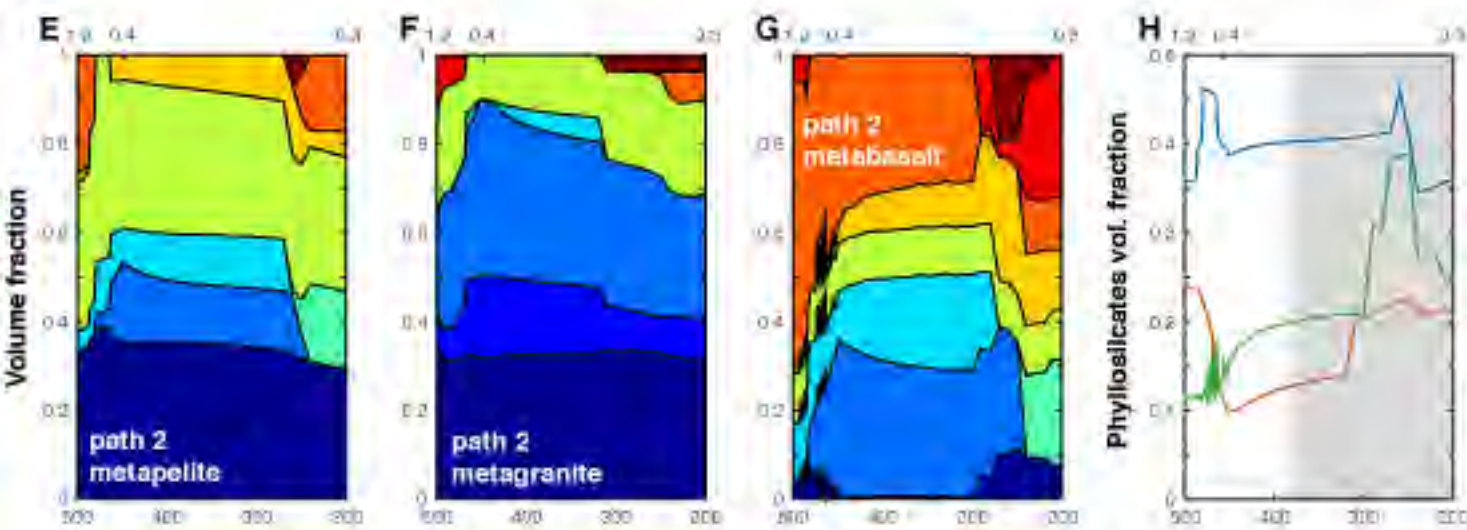




Pressure (GPa)



Temperature (°C)

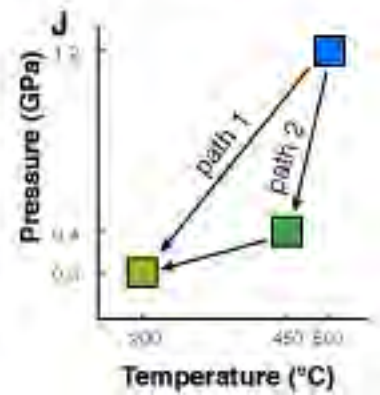


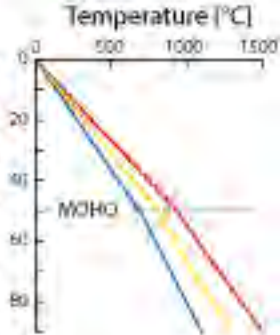
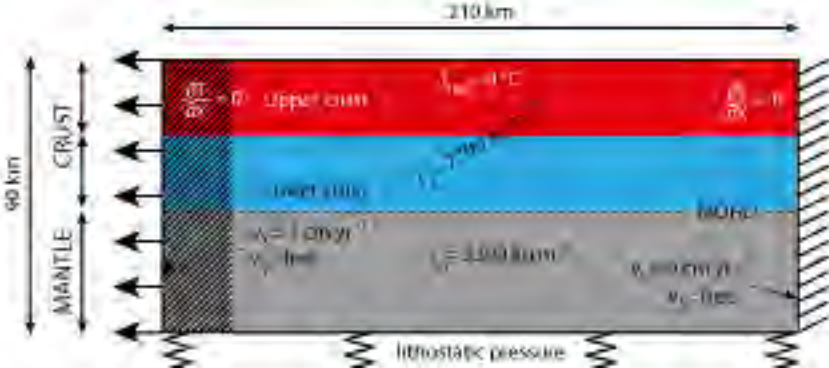
Temperature (°C)



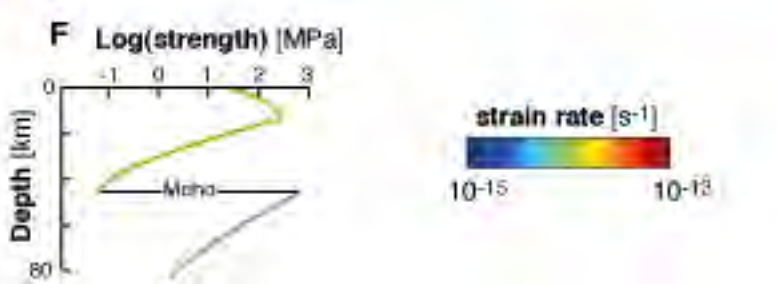
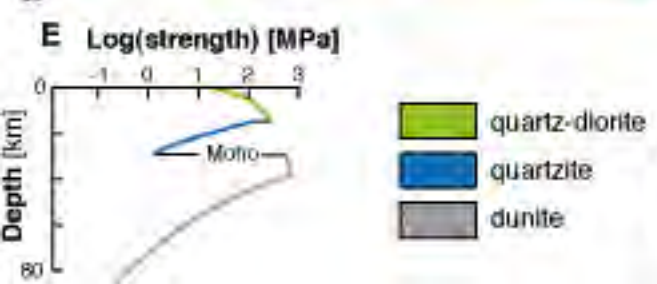
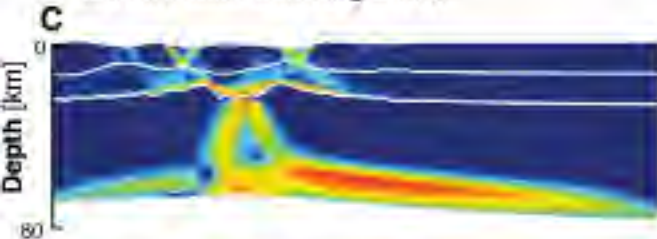
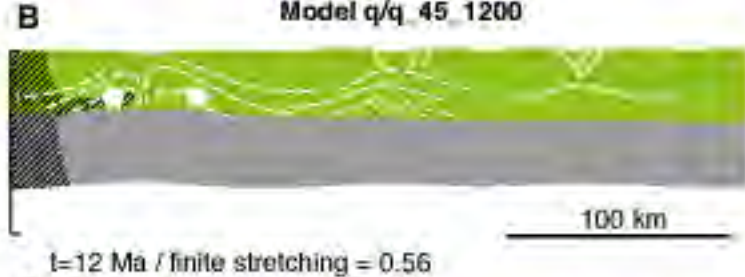
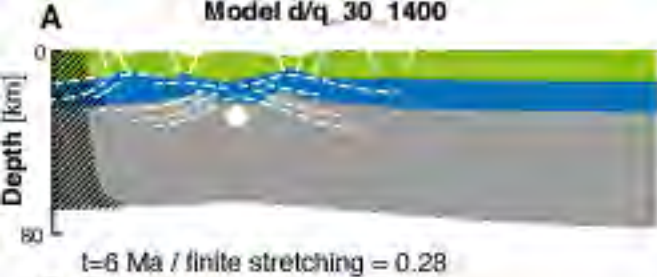
**I**

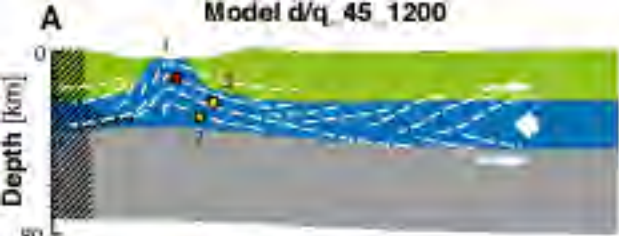
	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	FeO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O
Pelite (B95)	63.31	17.22	0.82	5.45	3.00	3.52	1.48	3.64
Granite (LM76)	71.84	14.43	1.22	1.65	0.72	1.85	3.71	4.00
Basalt (LM76)	49.97	15.99	3.85	7.00	8.84	9.82	2.96	1.12



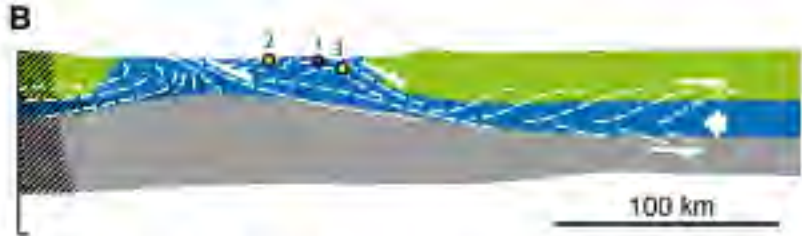
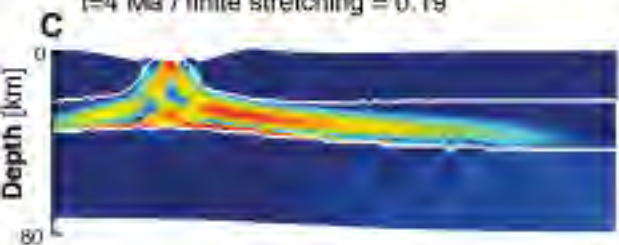




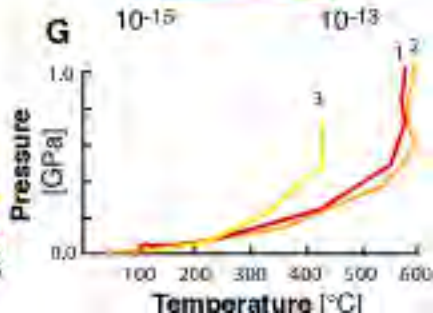
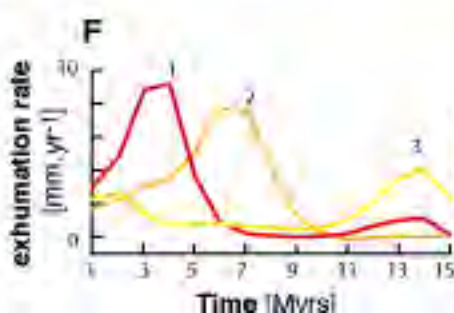
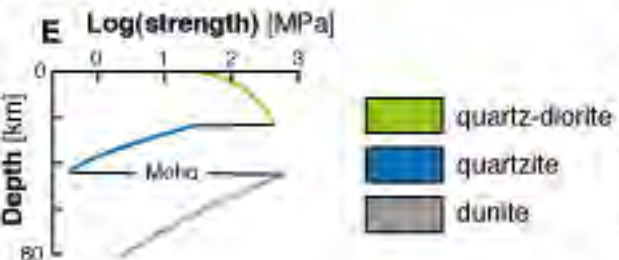
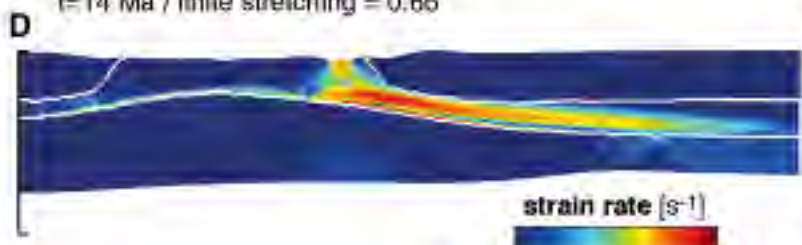


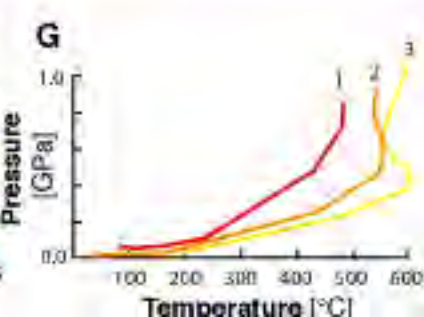
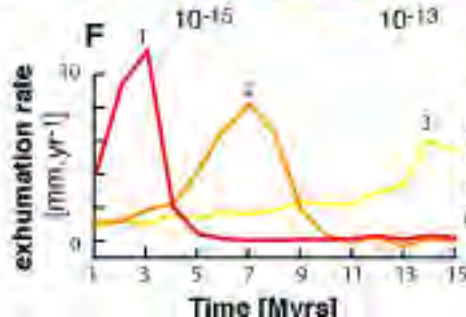
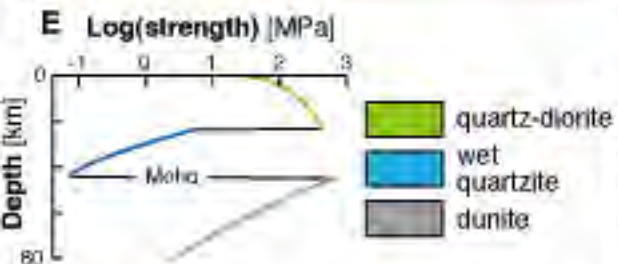
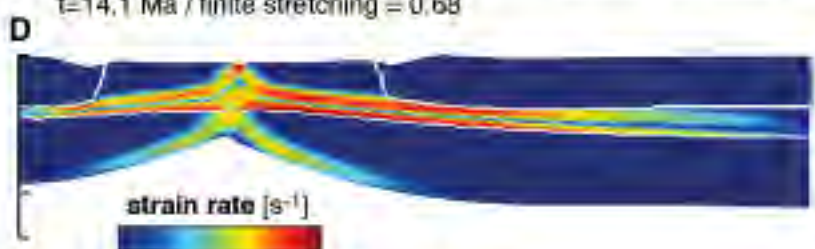
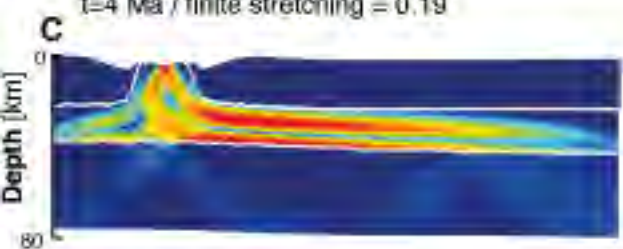
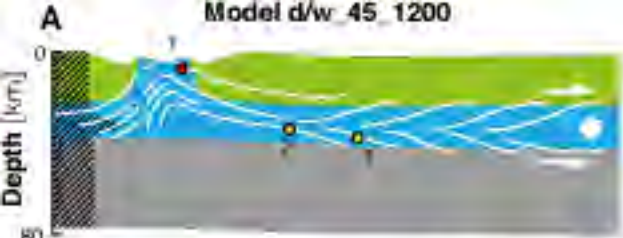


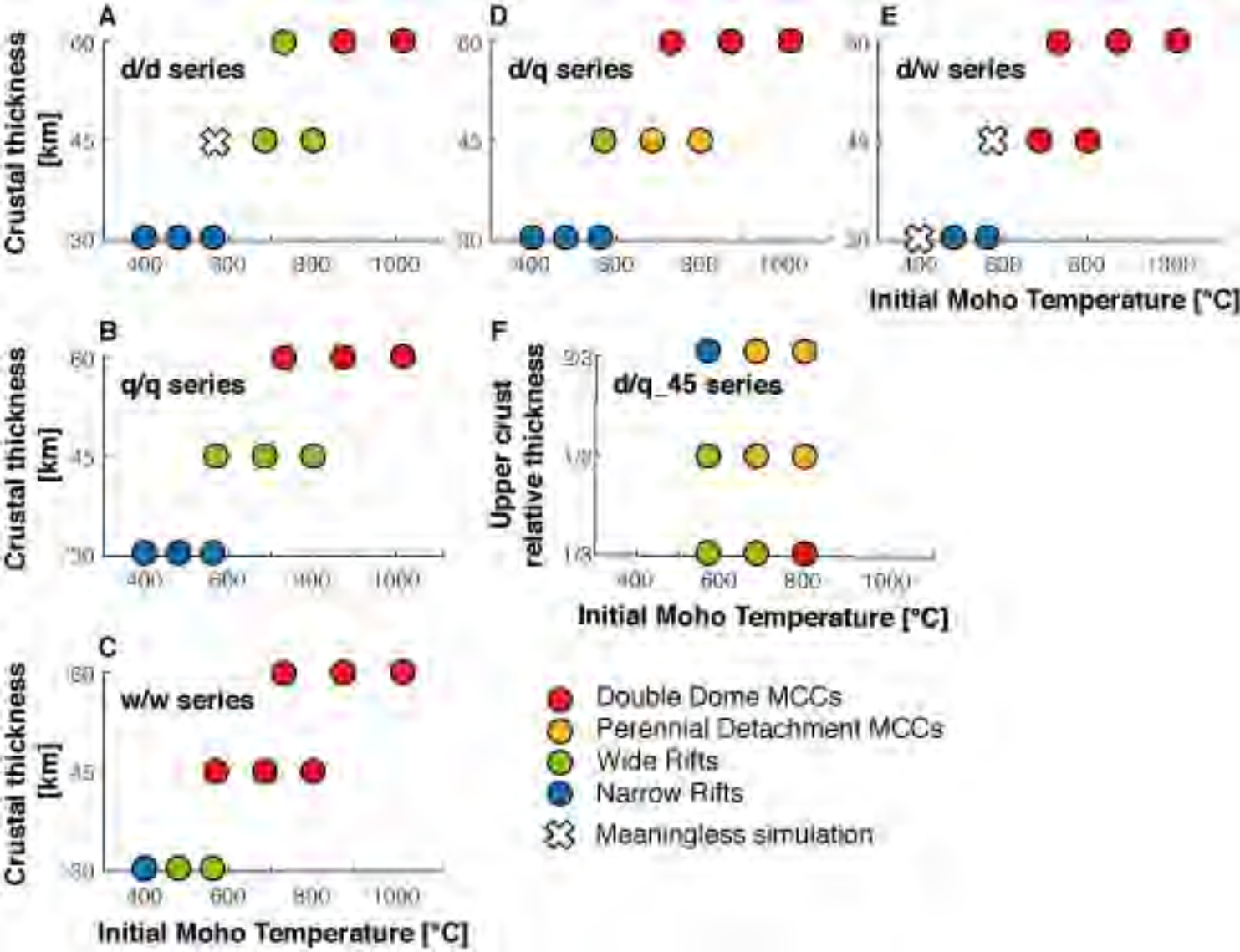
$t=4$  Ma / finite stretching = 0.19

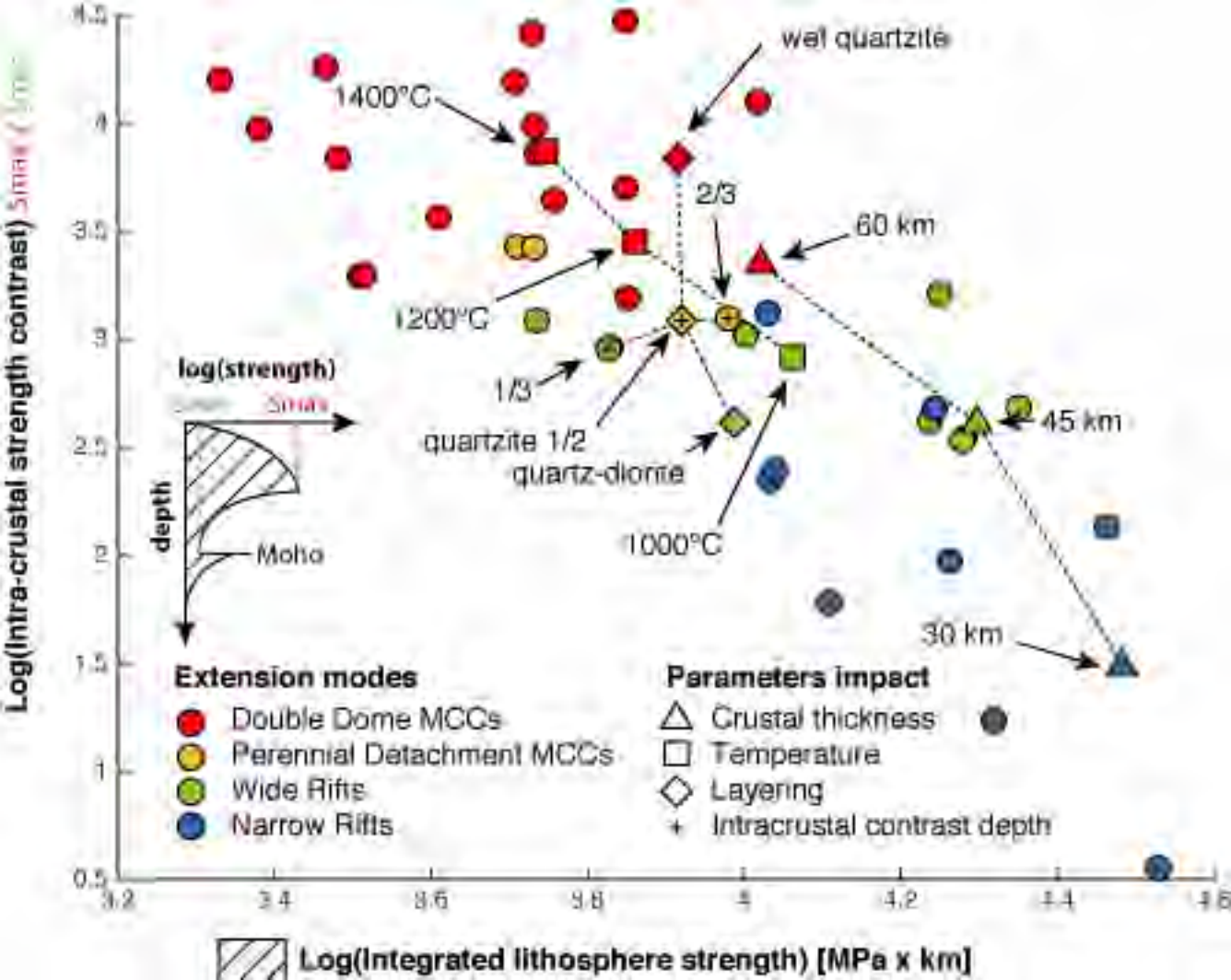


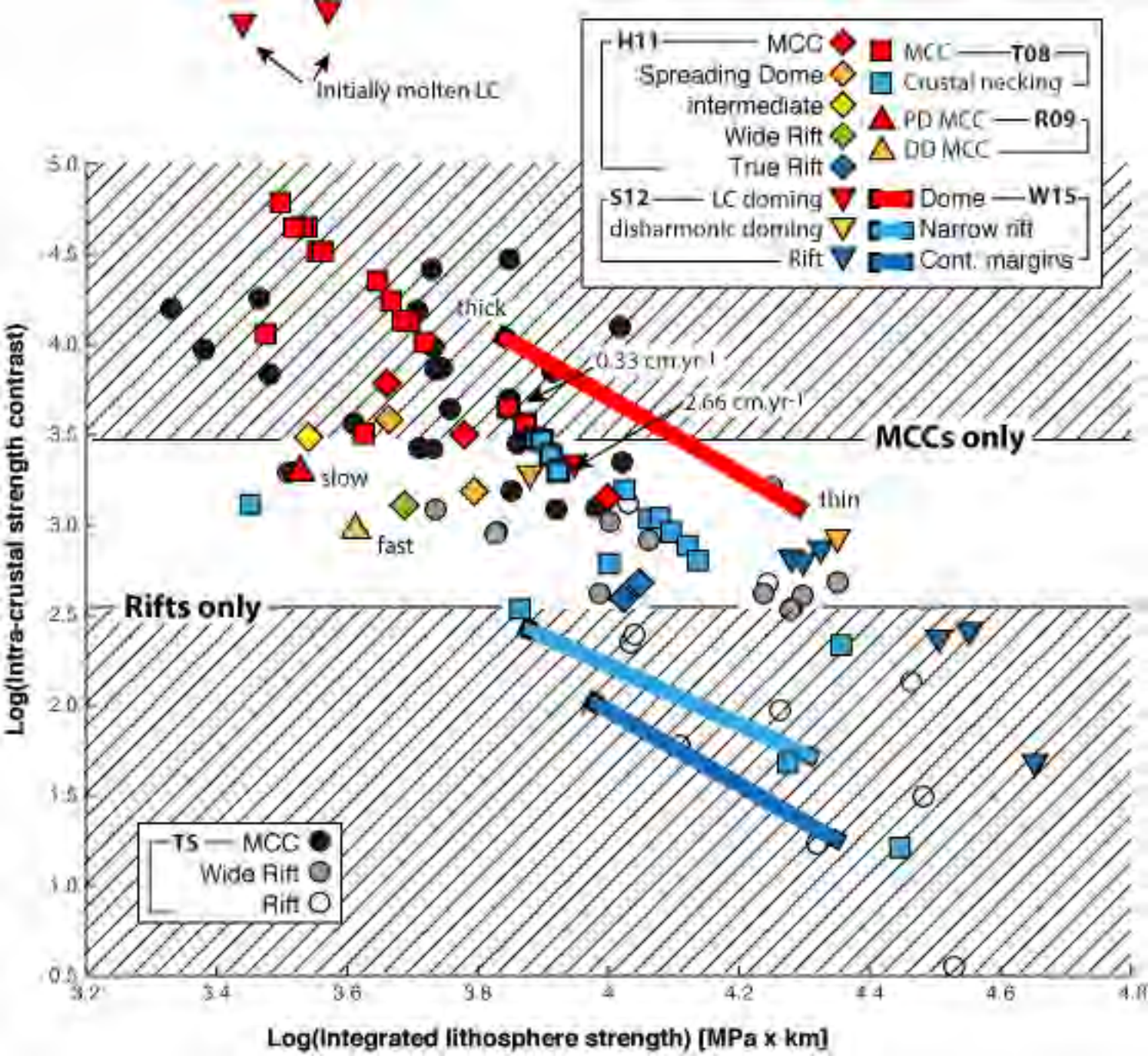
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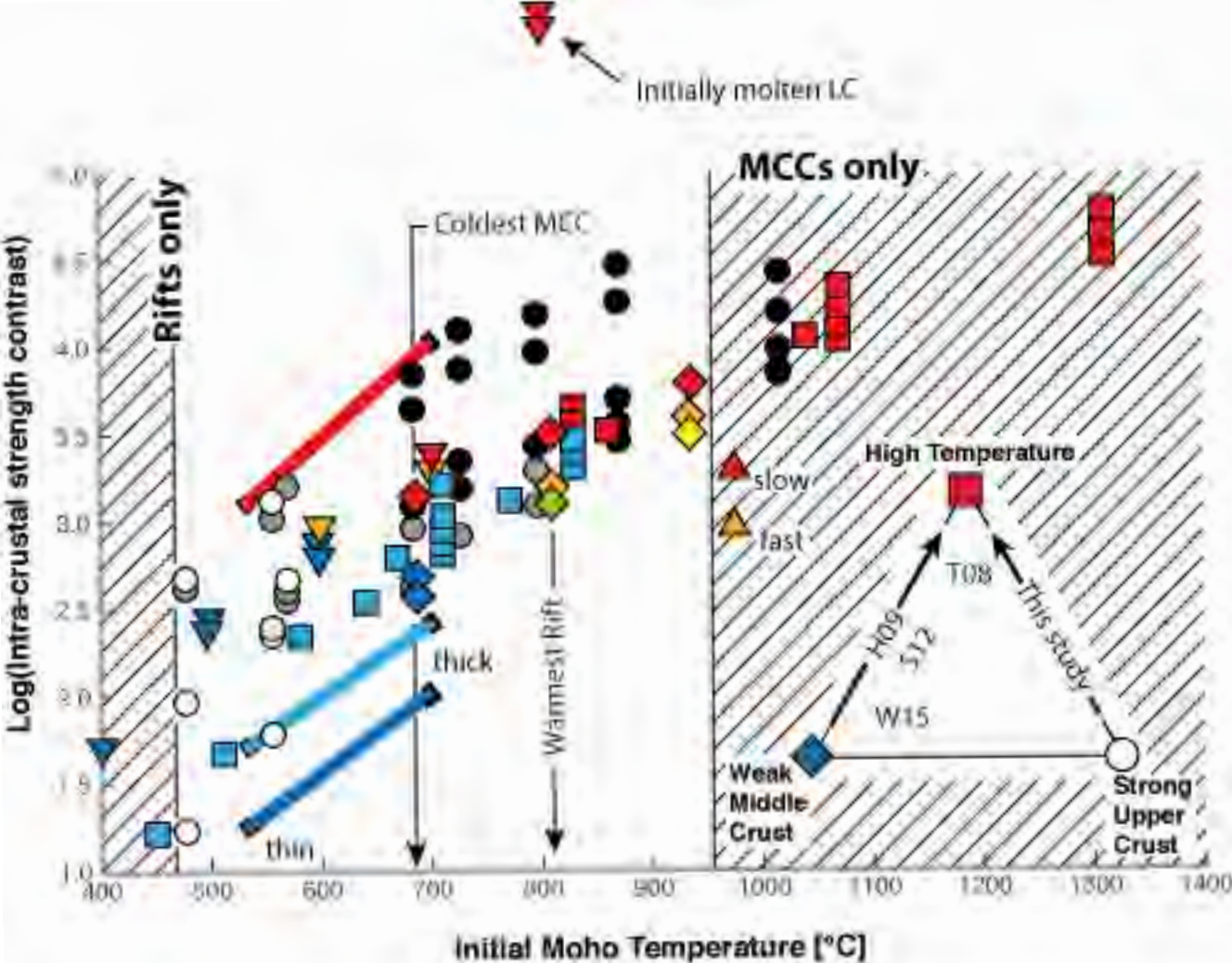


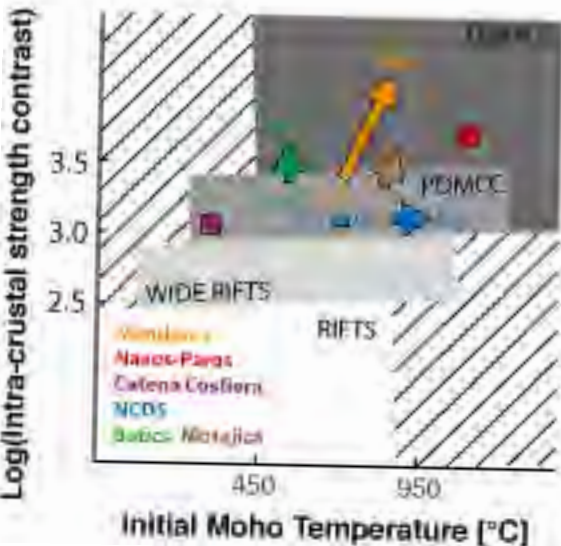




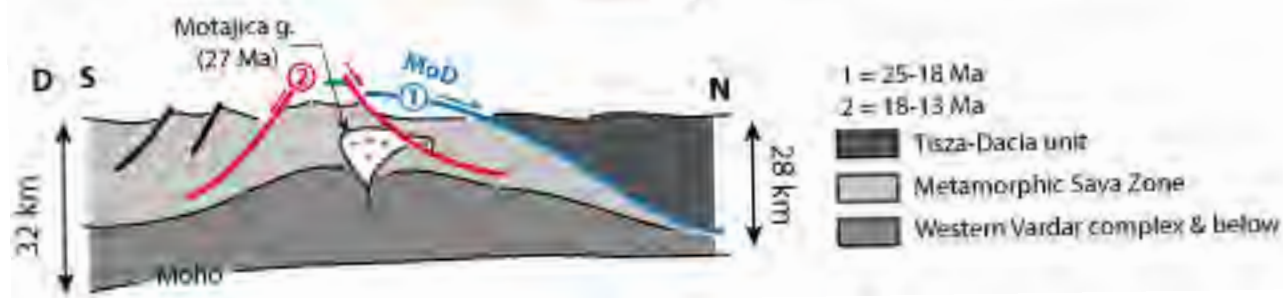
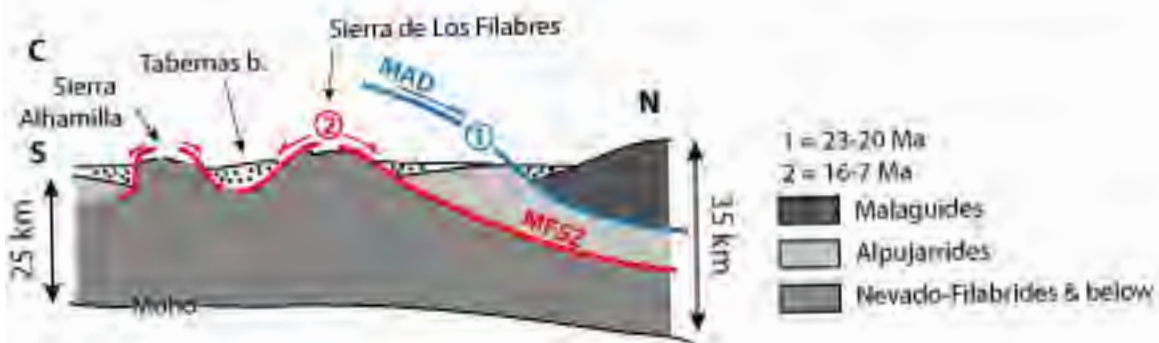
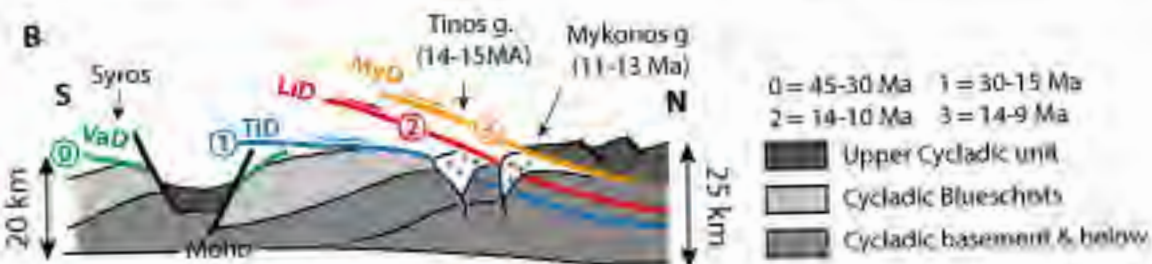


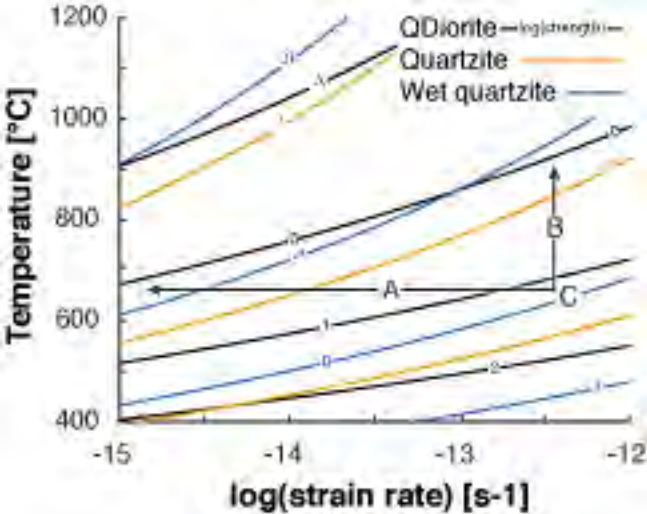












<b>Site</b>	<b>minimum pressure (kb)</b>	<b>maximum pressure (kb)</b>
SABN	8	10
GK	3	5
Edough	12	14
Tenda		
Beni Mesala	13	18
Filali		
BB kinzigites	9	13
BB mafic	16	20
Catena Costiera	11	13
Apuane Auto	7	9
Apuane Massa	8	11
argentario	11	13
aspromonte	11,2	12,4
alpu LG		7
alpu Es	6,5	8,5
alpu He	10	12
alpu S & A	7	11
NF BM	18	20
NF CA	11	15
NF beginning D3	3	4
Menderes Selimye	4	8
Menderes Bozdog		
Menderes Cine HP		
Menderes Cine HT		
Erhodope HP	13	15
Erhodope HT	3	9
Thasos	6,5	9,5
kazdag	4	6
Nigde Gumusler	5	6
andros post-orogen	6	8
Ikaria	6	8
tinos (transition)	8	10
tinos (late)	0	2
tinos HP	15	18
tinos BS	8	13
syros HP	16	20
syros retrogression	3	6
ios overprint		
ios peak basement	15	18
naxos	4,5	9,5
rechnitz HP	6	8
rechnitz late	3	4
veporic	9	10
Motajica	5	7

**minimum T (°C) maximum temperature (°C) average pressure (kb)**

660	700	
650	700	
600	800	
		9
450	550	
		8
800	850	
760	820	
	350	
410	430	
450	500	
350	400	
540	570	
	400	
	420	
540	620	
390	450	
550	650	
520	620	
		8
		15
		7
450	600	
550	620	
550	650	
590	690	
700		
600	650	
500	550	
400	500	
450	500	
375	525	
450	550	
300	450	
		8
480	520	
620	720	
330	370	
350	450	
550	630	

<b>average temperature (°C)</b>	<b>Code</b>	<b>Timing</b>	<b>ref</b>
	GK1	syn-detachment	saadallah96
	GK2	syn-detachment	"
	Ed	syn-detachment	caby01
400	Tn	syn-detachment	gueydan03
	BM	syn-detachment	negro 07
780	BB3	syn-detachment	"
	BB2	syn-detachment	"
	BB1	inherited	"
	CC	syn-detachment	rossetti04
	AA2	syn-detachment	jolivet98
	AA1	inherited	"
	Arg	syn-detachment	brunet00
	Asp	syn-detachment	ortolano05
	Alpu4	syn-detachment	azanon00
	Alpu3	inherited	"
	Alpu1	inherited	"
	Alpu2	inherited	"
	NF1	inherited	augier05
	NF2	inherited	"
500	NF3	syn-detachment	"
550	Me4	syn-detachment	whitney02 / Regnier 03
530	Me3	syn-detachment	okay01
640	Me1	inherited	oberhansli&al in press
620	Me2	syn-detachment	"
	Rho1	inherited	bonev & beccaletto07
	Rho2	syn-detachment	"
	Rho3	syn-detachment	brun & sokoutis07
	Kaz	syn-detachment	cavazza09
	Nig	syn-detachment	Whitney & Dilek, 98
450	And	syn-detachment	huet phd
		syn-detachment	Laurent 2016
	Tin3	syn-detachment	parra 02
	Tin4	syn-detachment	"
	Tin1	inherited	"
	Tin2	inherited	"
	Syr1	inherited	trotet 01
	Syr2	syn-detachment	huet phd
400	Ios1	syn-detachment	huet phd
	Ios2	inherited	huet phd
	Nax	syn-detachment	buick & holland 89
	Rech1	inherited	koller 85
	Rech2	syn-detachment	"
620	Vep	syn-detachment	janak 01
	Mot	syn-detachment	Ustaszewski10