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Timescales of subduction initiation and evolution of subduction thermal regimes

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2 thermal regimes

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37 **Highlights**

38 - Titanite and monazite in metamorphic soles record subduction initiation dynamics

39 - Initiation of the Neotethys subduction (Oman-UAE) was tectonically-induced prior to

40 102-100 Ma

41 - Subduction first stagnated for 5-10 Myr at warm conditions (15-20°C/km)

42 - Subduction evolved into a faster and colder self-sustained regime after 96-95 Ma

43

44 **Abstract**

45 Subduction zones are first-order features of plate tectonics on Earth, yet the mechanisms by
46 which subduction initiates remain enigmatic and controversial. Here, we reappraise the timing
47 of metamorphism of the rock units first detached from the leading edge of the downgoing slab
48 during initiation of the Neotethys subduction, now preserved in the metamorphic sole of the
49 Semail ophiolite (Oman–United Arab Emirates). Using petrochronology and phase equilibrium
50 modeling, we demonstrate that subduction initiated prior to 102–100 Ma at a slow rate (< 1
51 cm/yr). Subduction stagnated at relatively warm conditions ($15\text{--}20^\circ\text{C/km}$) for > 5 Myr before
52 evolving into a faster ($\geq 2\text{--}5 \text{ cm/yr}$) and colder ($\sim 7^\circ\text{C/km}$) self-sustained regime. Subduction
53 acceleration (i.e. “unlocking” stage) triggered the onset of slab retreat, large-scale corner flow
54 and fast ocean spreading in the overriding plate at 96–95 Ma, through the progressive change
55 of thermo-mechanical structure of the plate interface. This study reconciles conflicting
56 analogue and numerical subduction initiation models, shedding light on the thermal, mechanical
57 and kinematic complexity of subduction initiation.

58

59

60 **Keywords**

61 - Petrochronology

62 - Metamorphic sole

63 - Semail ophiolite

64 - Subduction initiation

65 - Thermal evolution

66

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68

69 **1. Introduction**

70 Subduction is one of the main facilitators of plate tectonics and yet the mechanisms responsible
71 for subduction initiation are controversial (Stern and Gerya, 2017). How plate-scale mechanics
72 and kinematics allow subduction to progress and reach steady-state remains unclear (Agard et
73 al., 2018; Anczkiewicz et al., 2004; Cramer et al., 2020; Gurnis et al., 2004; Holt and Condit,
74 2021). Direct geological records of subduction initiation are limited to metamorphic soles,
75 which correspond to thin sheets of deformed and metamorphosed oceanic crust. Metamorphic
76 soles underlie the mantle section of most ophiolitic complexes (e.g., Oman–United Arab
77 Emirates, Turkey, Newfoundland; Wakabayashi and Dilek, 2003) and are interpreted as
78 remnants of the tip of the downgoing slab metamorphosed and deformed at high temperature
79 along the nascent subduction interface (Agard et al., 2016; Wakabayashi and Dilek, 2003; Fig.
80 1a). These 10–500 m thick soles show remarkably similar characteristics worldwide (Agard et
81 al., 2016), with a downward decrease in metamorphic grade from granulite/amphibolite facies
82 to greenschist facies and an increase in the proportion of metasediments. This inverted sequence
83 attests to sequential removal of crustal slices from the slab (e.g. HTa and HTb in Fig. 1a) and
84 accretion to the base of the ophiolite during subduction initiation (Gnos, 1998; Soret et al.,
85 2017). Metamorphic soles therefore mark episodes of viscous coupling along the plate interface
86 before strain localization – and hence subduction – is achieved (Fig. 1b). With cooling of the
87 interface, the depth of viscous coupling between the downgoing plate (‘slab’) and the overlying
88 mantle is expected to migrate downwards (Agard et al., 2020) to 80–100 km in mature
89 subduction zones (Syracuse et al., 2010). This deepening of viscous coupling enables
90 subduction initiation (a mechanism coined "slabitization" by Agard et al., 2020) and may trigger
91 and control mantle upwelling and counterflow in the upper plate ophiolite (Fig. 1b).

92

93 Constraining the relative timing and duration of these coupling events is essential to
94 understanding both subduction initiation and the birth of ophiolites (Fig. 1a). Two conflicting
95 end-member models for subduction initiation have been proposed (Stern, 2004): (1) “induced”
96 or “far-field forced” subduction initiation, in which subduction nucleates in response to a
97 regional change in plate dynamics (Agard et al., 2007; Duretz et al., 2016; Gurnis et al., 2004;
98 Hall et al., 2003; Moores, 1982; Rodriguez et al., 2021; Van Hinsbergen et al., 2015), (2)
99 “spontaneous” subduction initiation, in which subduction ensues from the gravitational sinking
100 of dense lithosphere in the absence of significant plate convergence (Arculus et al., 2015;
101 Dymkova and Gerya, 2013; Stern, 2004). Regardless of the model, the upper plate ophiolite is
102 thought to form by upwelling of supra-subduction mantle once subduction has initiated (e.g.,
103 Rioux et al., 2016), and numerical and analogue experiments (Duretz et al., 2016; Stern and
104 Gerya, 2017) only reproduce subduction initiation with fast-paced convergence ($> \text{cm/yr}$) along
105 a pre-existing lithospheric-scale weak zone.

106

107 $^{40}\text{Ar}/^{39}\text{Ar}$ and U–Pb geochronology of metamorphic soles and associated ophiolites (Garber et
108 al., 2020; Gnos and Peters, 1993; Hacker, 1994; Hacker et al., 1996; Rioux et al., 2021b, 2016,
109 2013; Soret et al., 2016) commonly advocate for synchronous slab burial and ocean spreading
110 in the upper plate, as predicted by the spontaneous subduction initiation model (Stern, 2004).
111 However, garnet Lu–Hf dating of the highest-grade sole unit (accreted first, HTa on Fig. 1a, b)
112 in the Semail and Turkish exposures has recently revealed burial to granulite facies ($\sim 850^\circ\text{C}$)
113 ~ 8 Myr earlier than previously estimated, i.e. 103–104 Ma (Guilmette et al., 2018; Pourteau et
114 al., 2018) instead of 96–95 Ma (Gnos and Peters, 1993; Hacker, 1994; Hacker et al., 1996;
115 Rioux et al., 2016). U–Pb dating of zircon rims included in garnet from a lower-grade unit (HTb;
116 $T \sim 650^\circ\text{C}$) confirms that prograde metamorphism started $\geq 98.7 \pm 1.7$ Ma (Garber et al., 2020).
117 Dates from the cores of the same zircon grains ($107\text{--}102 \pm 2$ Ma), though coeval within

118 uncertainty with the garnet dates from the HTa unit (Guilmette et al., 2018), were interpreted
119 as detrital (Garber et al., 2020).

120

121 These few garnet and zircon dates from the HTa and HTb units indicate that subduction initiated
122 before the formation of the associated supra-subduction zone ophiolite (96–95 Ma; Rioux et
123 al., 2016), supporting the view that ophiolite-sole couples record far-field forcing before upper
124 plate extension (Agard et al., 2007; Guilmette et al., 2018; Rodriguez et al., 2021; Van
125 Hinsbergen et al., 2015) rather than spontaneous subduction initiation. Yet, these older dates
126 have not yet been robustly ascribed to specific P–T conditions and as such provide only little
127 information on the timescale and rate of slab sinking during subduction initiation (Fig. 1b).

128

129 This study uses thermodynamic modeling tied to *in situ* titanite and monazite dating of the sub-
130 units of the Semail metamorphic sole (Oman–UAE) to precisely quantify the temporal record
131 of their burial, accretion and exhumation. This new dataset provides direct information on i)
132 the onset, duration and rate of subduction initiation, ii) thermo-mechanical processes along the
133 plate interface and iii) the link between these processes and the supra-subduction ophiolite
134 formation.

135

136 **2. The Semail metamorphic sole units**

137 The Semail metamorphic sole comprises a dominantly mafic, high-temperature (HT) unit thrust
138 onto a mainly metasedimentary low-temperature (LT) unit (Fig. 2a, b; Gnos, 1998). The HT
139 sole itself consists of two tectonic slices juxtaposed sequentially, HTa then HTb (Soret et al.,
140 2017). The latter is capped by a meter-thick garnet-rich metasedimentary layer. Specimens were
141 collected from the Khubakhib and Sumeini half-windows (UAE and Oman, respectively; Fig.

142 2a, b), where the structure of the Semail metamorphic sole is well preserved (Gnos and Peters,
143 1993; Soret et al., 2017).

144

145 **2.1. Mineralogy and P-T conditions of formation**

146 The strongly foliated granulite-facies metabasalts of HTa preserve a prograde to peak
147 metamorphic assemblage of garnet, clinopyroxene, amphibole and minor plagioclase (Cowan
148 et al., 2014; Gnos, 1998; Guilmette et al., 2018; Soret et al., 2019a, 2017), despite partial
149 overprinting at amphibolite facies conditions. Titanite occurs as euhedral to subhedral
150 inclusions in both prograde and peak phases (i.e. garnet and clinopyroxene porphyroclasts; Ti-
151 rich amphibole; Fig 3a-c), forming locally rotated inclusion trails (Gnos, 1998). Rutile and
152 zircon occur as small inclusions (1–10 μm) in garnet and titanite (Fig. 3b).

153

154 Amphibolite-facies metabasalts of HTb mainly consist of amphibole and plagioclase with
155 minor biotite and either clinopyroxene or chemically zoned garnet (Garber et al., 2020; Soret
156 et al., 2019a, 2017). Titanite forms euhedral to subhedral grains closely associated with both
157 amphibole and plagioclase, either as inclusions or along boundaries (Fig. 3e-f). Rutile and
158 zircon form small inclusions (1–30 μm) in titanite and amphibole (Fig. 3e). In the presence of
159 garnet, ilmenite is observed instead of titanite (Soret et al., 2017).

160

161 The sheared metasedimentary layer at the top of HTb is dominated by quartz, garnet, muscovite
162 and plagioclase. Garnet cores host quartz and ilmenite inclusions (Fig. 3g). Garnet mantles
163 contain minute zircon grains as well as abundant polycrystalline inclusions with a negative
164 crystal shape (Fig. 3g-i), indicating peritectic growth reaction (Cesare et al., 2009; Dubacq et
165 al., 2019). Garnet rims are thin ($\leq 150 \mu\text{m}$) and typically inclusion-free, save for minute

166 monazite and zircon (1–10 μm). Large monazite (50–100 μm ; Fig. 3j, k) occurs in the rock
167 matrix where it is rimmed by a syn-kinematic assemblage of epidote, allanite and apatite.

168

169 Peak P–T conditions for the HTa, HTb and LT units of the sole are estimated at $850 \pm 50^\circ\text{C}$
170 and 1.0 ± 0.1 GPa, $725 \pm 50^\circ\text{C}$ and 0.8 ± 0.1 GPa, and $530 \pm 50^\circ\text{C}$ and 0.5 ± 0.1 GPa,
171 respectively (Soret et al., 2017). For HTa, no significant petrological/P-T variations have been
172 found along strike, despite being separated by hundreds of kilometers beneath the ophiolite
173 (Khubakhib, Sumeini and Wadi Tayin areas; Soret et al., 2017). The P–T conditions for HTb
174 are less precisely known (Soret et al., 2017; Garber et al., 2020) and therefore reassessed here
175 (see section 4.1).

176

177 **2.2. Existing geochronology**

178 Zircon U–Pb dating of melt segregations in HTa indicates melt crystallization and incipient
179 cooling by 96.2–94.5 Ma (Guilmette et al., 2018; Rioux et al., 2016). Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$
180 dating indicates cooling below 500–550°C (Ar diffusion closure temperature) shortly
181 thereafter, between 95.7 and 92.6 Ma (Gnos and Peters, 1993; Hacker et al., 1996). A subset of
182 four titanite grains in HTa yielded a single $^{206}\text{Pb}/^{238}\text{U}$ age of 95.6 ± 0.3 Ma, which was
183 interpreted as a cooling age below 600–650°C (Guilmette et al., 2018). Finally, titanite and
184 monazite U–Pb dating combined with garnet Lu–Hf dating in a lower grade unit (peak
185 $T \sim 650^\circ\text{C}$) indicate prograde to peak growth age from 98.7 ± 1.7 Ma to 93.0 ± 0.5 Ma (Garber
186 et al., 2020). A single titanite U–Pb date of 92.2 ± 1.8 Ma is interpreted to record the initial
187 cooling of this unit (Garber et al., 2020). Fast cooling of both HTa and HTb sole units shortly
188 after reaching peak conditions is further supported by the preservation of prograde chemical
189 zoning in garnet (Garber et al., 2020; Guilmette et al., 2018; Soret et al., 2019a, 2017).

190

191 **3. Results**

192 A detailed description of the different methods used in this study is provided in the
193 supplementary materials.

194 **3.1. Phase equilibrium modeling**

195 The P–T evolutions of the two garnet-bearing HTb specimens (metasediment: SE13-123L;
196 metabasalt: SE16-26) determined through forward thermodynamic modeling (Perple_X
197 software version 6.8.6; Connolly, 2009; Connolly and Kerrick, 1987) are shown in Figure 4a-
198 c. The P–T history of the HTa unit has previously been estimated by Soret et al. (2017). Mineral
199 abbreviations throughout the text, figures, tables and supplementary information are after
200 Whitney and Evans (2010).

201
202 The pseudosections (Fig. 4a-b) for mafic composition from the HTb unit (Tab. SM3) have been
203 modeled in the NCKFMASHT system using the internally consistent thermodynamic dataset
204 (hp62ver.dat) of Holland and Powell (2011). The H₂O value was set to 1.0 wt.% after
205 calculating a T-X(H₂O) diagram (Fig. 3a) with a water content ranging from 0.5 wt.% to 2.0
206 wt.% (i.e., excess water) at a pressure of 0.75 GPa consistent with the estimates from the
207 amphibole-plagioclase thermo-barometry (~0.75 GPa and 750°C; inset of Fig. 3a; Holland and
208 Blundy, 1994; Molina et al., 2015). The H₂O value is similar to that reported in the literature
209 for similar metamorphic sole amphibolites (Mont Albert, Andaman, Oman-UAE; Dubacq et
210 al., 2019; Plunder et al., 2020; Soret et al., 2017). The pseudosection (Fig. 4c) for the felsic
211 composition (Tab. SM3) has been modelled in the CKFMASH system using the internally
212 consistent thermodynamic dataset (hp02ver.dat) of Holland and Powell (1998). The H₂O
213 content was calculated to be in excess at subsolidus conditions and with a water-saturated
214 solidus at 0.8 GPa. The P–T estimates were determined based on the observed mineral
215 assemblage and the isopleth compositions of garnet (Fig. 4b, c) for both specimens because

216 most of the plagioclase, biotite and muscovite (re-)crystallized during retrograde
217 metamorphism (Soret et al., 2017).

218

219 Peak P–T conditions inferred for the Grt + Amp + Pl + Bt + Ilm ± Qz assemblage of the garnet-
220 bearing amphibolite (SE16-26, Sumeini; Fig. 4a, b) lie at sub-solidus conditions of $725 \pm 25^\circ\text{C}$
221 and 0.77 ± 0.03 GPa. In addition to the calculated garnet rim composition (Alm₆₅₋₆₆Prp₁₇₋
222 ₁₈Grs₁₇), calculated modes (Grt₁₀Amp₃₅Pl₃₈Bt₁₀Ilm_{2.5}) and plagioclase composition (An₄₁) are
223 in close agreement with those observed in the specimen (Soret et al., 2017). Garnet core
224 isopleths converge at $625 \pm 20^\circ\text{C}$ and 0.70 ± 0.02 GPa in the same stability field.

225

226 Peak P–T conditions for the garnet-bearing metasediment (SE13-123L, Khubakhib; Fig. 4c) lie
227 above the wet solidus for felsic rocks, at $750 \pm 10^\circ\text{C}$ and 0.76 ± 0.02 GPa, in the field of Grt +
228 Bt + Qz + Sil + Pl + WM + melt. These conditions were determined using isopleth intersections
229 for the peritectic garnet mantle (Alm₆₉₋₇₁Prp₁₈₋₁₉Grs₁₂₋₁₃). The composition of the quartz-rich
230 garnet core (Alm₆₇₋₆₈Prp₁₃₋₁₄Grs₁₈₋₁₉) indicates a prograde, near-isobaric evolution from $610 \pm$
231 10°C and 0.72 ± 0.2 GPa. The thin inclusion-free garnet rim (Alm₇₆₋₇₇Prp₁₂Grs₁₁) further
232 indicates sub-solidus re-equilibration during retrograde P–T conditions at $660 \pm 20^\circ\text{C}$ and 0.61
233 ± 0.03 GPa, in the white mica stability field. Considering the low proportion of sillimanite (< 9
234 vol%) expected along the PT path, its absence in the specimen could be explained by the
235 extensive LT retrogression (Fig. 3g) and/or by reaction overstepping due to garnet nucleation
236 (Spear and Pattison, 2017).

237

238 **3.2. *In situ* titanite petrochronology**

239 *In situ* U–Th/Pb and trace-element analyses of titanite and monazite (see section 3.3) were
240 carried out on five HTa specimens and five HTb specimens (Fig. SM2, SM3) using the LASS

241 facility at the University of California, Santa Barbara (USA). Data reduction for U–Th/Pb,
242 including corrections for baseline, instrumental drift, mass bias, down-hole fractionation were
243 carried out using Iolite v. 2.5 (Paton et al., 2011), and error correlations were recalculated after
244 Schmitz and Schoene (2007). Methods used in this study follow those outlined by Kylander-
245 Clark et al. (2013), with modifications presented by McKinney et al. (2015).

246

247 Rutile and zircon, though present in all specimens, proved too small ($\leq 20 \mu\text{m}$) to be analyzed.
248 Different titanite age populations within each specimen in both HTa and HTb units were
249 distinguished based on their textural position and Rare Earth elements (REE) contents (Fig. 5).
250 Titanite ages were calculated for all titanite populations. They correspond to ^{207}Pb -corrected
251 (following Stacey and Kramers, 1975) $^{206}\text{Pb}/^{238}\text{U}$ weighted means, and they ranged between
252 94.7 and 100 Ma with MSWD between 0.9 and 2.7 (Tab. 1).

253

254 3.2.1. *The HTa metamorphic sole unit*

255 In SE11-23 (Sumeini), titanite grains are included in the core (G1) and the rim (G2) of a garnet
256 porphyroclast (Fig. 3c, 5a). Both inclusion sets show a depletion in HREE, while G2 is enriched
257 in LREE and Zr compared to G1 (Fig. 5a-c). G1 grains yielded older date ($99.5 \pm 0.7 \text{ Ma}$; spot
258 analyses ($n = 27$) than those found at the rim (G2; $97.2 \pm 0.3 \text{ Ma}$, $n = 42$).

259

260 In SE14-31 (Sumeini), titanite grains are either included at the core of a cm-scale garnet and a
261 mm-scale clinopyroxene or in the matrix. Five titanite populations were distinguished based on
262 variations in the textural position and REE concentrations (Fig. 5a, 5d-h). The first population
263 (G1) corresponds to inclusions in the garnet core. These titanite have flat REE patterns and
264 returned the oldest date ($100.0 \pm 2.0 \text{ Ma}$; $n = 14$). The second population (G2) comprises
265 inclusions in the core of both garnet and clinopyroxene. It exhibits a slight HREE depletion and

266 enrichment in Zr relative to G1, and yielded a date of 98.0 ± 0.9 Ma ($n = 35$). The third
267 population (G3) also comprises inclusions in the core of garnet and clinopyroxene, but has a
268 distinct REE pattern marked by an enrichment in LREE, a larger HREE depletion, and higher
269 Zr concentrations than G2. G3 titanite yielded a date of 95.8 ± 0.5 Ma ($n = 74$). The fourth
270 titanite population (G4) in this specimen consists of inclusions in the core of garnet and
271 clinopyroxene together with titanite found in the matrix. These grains show a higher enrichment
272 in LREE with a similar HREE depletion, and the highest Zr concentrations of any of the titanite
273 populations in SE14-31. G4 titanite yielded a date of 95.9 ± 0.5 Ma ($n = 93$), which overlaps
274 with that of G3 grains. The last population (G5) includes solely titanite grains from the matrix.
275 The grains are characterized by a LREE content similar to G4 but with a significant enrichment
276 in HREE. G5 titanite returned the youngest date in the specimen (94.9 ± 0.7 Ma; $n = 35$).

277
278 In SE13-123A (Khubakhib), titanite grains are included at the rim of three slightly zoned garnet
279 porphyroclasts (Fig. 5a). These grains have a significant depletion in HREE relative to other
280 REE (Fig. 5i), and yielded a date of 96.9 ± 0.4 Ma ($n = 98$). In SE13-123B (Khubakhib), titanite
281 grains are found only in the matrix. They have high REE and Zr contents with a pattern marked
282 by a slight depletion in HREE relative to LREE (Fig. 5j). The titanite in this specimen returned
283 a date of 96.6 ± 0.9 Ma ($n = 57$). In SE13-123G (Khubakhib), titanite grains are included at the
284 core of a cm-scale garnet (Fig. 5a). They display a REE pattern characterized by a large
285 depletion in LREE relative to chondrites (Fig. 5k), and yielded a date of 99.6 ± 1.1 Ma ($n = 33$).

286
287 Titanite from the HTa unit has an average Zr content varying between 210 and 575 (± 50) ppm
288 which is negatively correlated with the calculated ^{207}Pb -corrected $^{206}\text{Pb}/^{238}\text{U}$ mean dates (Fig.
289 5b-k; Tab. 1).

290

291 3.2.2. *The HTb metamorphic sole unit*

292 At Sumeini, titanite is texturally associated with amphibole and plagioclase in garnet-free rocks.
293 In all specimens, it shows a slight depletion in HREE relative to the LREE (Fig. 5l-r). Different
294 age populations within each specimen were distinguished based on variations in Zr and REE
295 concentrations. In SE14-16B (Fig. 5l-m), G1 titanite yielded a date of 97.2 ± 1.4 Ma ($n = 15$),
296 while G2 returned a date of 96.3 ± 0.8 Ma ($n = 36$). In SE14-17C (Fig. 5n-o), G1 gave a date
297 of 98.8 ± 1.1 Ma ($n = 16$) and G2 yielded a date of 97.0 ± 0.6 Ma ($n = 31$). In SE16-20 (Fig.
298 5p-q), G1 returned a date of 95.2 ± 1.3 Ma ($n = 9$), while G2 gave a date of 95.4 ± 0.8 Ma ($n =$
299 21). In SE16-11 (Fig. 5r), titanite grains are homogenous in composition and yielded a date of
300 94.7 ± 0.8 Ma ($n = 23$).

301

302 The different titanite populations from the HTb unit have an average Zr content varying
303 between 150 and 310 (± 25) ppm, with the highest concentration in the 96-95 Ma old population
304 (Fig. 5l-r; Tab. 1).

305

306 **3.3. *In situ* monazite petrochronology**

307 Eighteen monazite grains were dated in a metasediment sampled at the top of the HTb unit in
308 Khubakhib (Fig. 3j-k, SM3). All grains are located in the matrix. Monazite (1–10 μm)
309 inclusions in garnet were too small to get a reliable laser ablation signal. 190 spot analyses were
310 collected across 18 grains yielding most $^{208}\text{Pb}^*/^{232}\text{Th}$ dates between 102 Ma and 96 Ma (Fig.
311 6a-b).

312

313 Two monazite age populations are recognized based on their REE patterns (Fig. 6c-d). The first
314 population (G1) contains high Gd/Yb ratios (2180–2670) and low Y content (260–320 ppm).

315 Three grains (including the core of a zoned grain) were large enough for multiple analyses and

316 yielded weighted mean $^{208}\text{Pb}^*/^{232}\text{Th}$ dates of 99.2 ± 0.5 Ma ($n = 27$, SE13-123L2, Z1), $99.5 \pm$
317 0.6 Ma ($n = 23$, SE13-123L2, Z9) and 98.5 ± 0.4 Ma ($n = 48$, SE13-123L2, Z10 core). The
318 second population is characterized by low Gd/Yb ratios (20-40) and high Y content (6920-
319 21870 ppm). Analyses from four grains yielded weighted means of 97.6 ± 1.2 Ma ($n = 7$, SE13-
320 123L1, Z9), 96.6 ± 1.3 Ma ($n = 7$, SE13-123L2, Z3), 97.8 ± 1.1 Ma ($n = 9$, SE13-123L2, Z8)
321 and 97.5 ± 1.5 Ma ($n = 6$, SE13-123L2, Z10 rim).

322

323 **4. Discussion**

324 **4.1. P–T–t evolution of the Semail metamorphic sole**

325 Phase equilibrium modeling of both mafic and felsic lithologies helps refine the metamorphic
326 evolution of the HTb unit, particularly the pressure history. Peak P–T conditions of $750 \pm 25^\circ\text{C}$
327 and 0.77 ± 0.03 GPa confirm the existence of a metamorphic gap with HTa ($\Delta T \sim 100^\circ\text{C}$, ΔP
328 ~ 0.2 GPa; Soret et al., 2017). The striking similarity of the clockwise P–T paths of HTb and
329 HTa (Fig. 7c) indicates a similar thermal history, characterized by a sub-isobaric temperature
330 increase ($>100^\circ\text{C}$).

331

332 *In situ* titanite and monazite petrochronology allows assessment of the P–T–time paths of HTa
333 and HTb (Fig. 7c). The trace element zoning in titanite (Fig. SM4) rules out significant diffusive
334 homogenization subsequent to crystallization, as empirically demonstrated at temperatures of
335 850°C (Holder et al., 2019). Zirconium-in-titanite thermometry (Hayden et al., 2008) results
336 are, therefore, interpreted to reflect crystallization temperatures (Fig. 7a).

337

338 In HTa, the negative slope in REE of titanite inclusions in the core of chemically zoned garnet
339 and clinopyroxene porphyroclasts (Fig. 5) confirms they crystallized in equilibrium with garnet.
340 The Zr-poor titanite population (200–300 ppm) yields temperature estimates ($750\text{--}800^\circ\text{C}$, at P

341 ~0.7–1.0 GPa) similar to those predicted by phase equilibrium modeling for garnet nucleation
342 at or near solidus conditions (Soret et al., 2017). Therefore, the weighted mean dates at 100–97
343 Ma are interpreted as the timing of garnet growth during increasing P-T conditions (time 1, t1;
344 Fig. 7c). The flat REE pattern and the low Zr content displayed by the oldest age population
345 (~100 Ma; G1 in SE14-31; Fig. 5c) could be interpreted as crystallization prior to garnet
346 nucleation in that particular specimen. Moreover, temperatures of 815–840°C (at P ~1.0 GPa)
347 are calculated from the Zr-rich titanite population (400–600 ppm). Because temperature
348 estimates of this population overlap the calculated supra-solidus peak metamorphic conditions
349 (Soret et al., 2017), the 96–95 Ma dates are interpreted as the age of peak P-T conditions (time
350 2, t2; Fig. 7c). The peritectic growth of this population is further supported by the enrichment
351 in LREE (Fig. 5c,f,g,i,j) relatively to the oldest populations (100–98 Ma; Fig. 5b,d,e,k). It
352 should also be noted that the earliest stages of garnet nucleation (> 100 Ma) recorded in some
353 garnet cores were associated with the co-crystallization of rutile instead of titanite (Fig.5a; Soret
354 et al., 2017), and given the relatively low closure temperature of rutile (Bonnet et al., *in press.*),
355 could not be constrained in time

356

357 In HTb, the average Zr content in titanite is systematically lower than in HTa at a given age,
358 consistent with cooler metamorphic conditions. The relatively flat REE pattern (Fig. 5) is also
359 consistent with crystallization in a garnet-free environment. Zirconium-in-titanite thermometry
360 yields a maximum temperature of 740–760°C (at P ~0.7 GPa), in agreement with peak
361 metamorphic conditions estimated for this unit (Fig. 4a-c). Related weighted mean dates of 96-
362 95 Ma are therefore interpreted as the time when HTb was accreted at peak conditions (t2; Fig.
363 7c). Older titanite grains (~99 Ma) contain less Zr on average, suggesting cooler temperatures
364 ~725°C (at P ~0.7 GPa), consistent with crystallization along the prograde path near peak
365 conditions and a lack of further re-equilibration of Zr.

366

367 Lead diffusivity in monazite is thought to be negligible up to temperatures of $>800^{\circ}\text{C}$ for grains
368 larger than $20\ \mu\text{m}$ and over a period of ~ 10 Myr (Cherniak et al., 2004). Monazite dates in HTb
369 are, therefore, interpreted to represent crystallization ages. The combination of high (Gd/Yb)
370 ratios and low Y content in the (102–)100–98 Ma monazite (G1; Fig. 7b) is consistent with
371 prograde co-crystallization with xenotime and/or garnet. This is in contrast with low (Gd/Yb)
372 ratios and high Y content of 98–96 Ma monazite (G2; Fig. 7b), which are compatible with either
373 xenotime breakdown at amphibolite facies or garnet breakdown at supra-solidus conditions
374 along the prograde P-T path (Shrestha et al., 2019; Soret et al., 2019b; Spear and Pyle, 2010)
375 (t2; Fig. 7c). Because i) monazite tends to dissolve upon partial melting (Yakymchuk and
376 Brown, 2014), ii) phase equilibrium modeling does not predict any garnet breakdown along the
377 prograde P-T path, and iii) monazite appears to be partly dissolved at its edge (Fig. 3j, k), we
378 suggest that monazite records the prograde evolution of the HTb unit at amphibolite-facies from
379 (102–)100 Ma to 96 Ma, consistent with titanite petrochronology.

380

381 **4.2. Kinematics, mechanics and thermal structure of subduction initiation**

382 The results of this study indicate that the leading edge of the nascent slab preserved in the HTa
383 and HTb metamorphic sole units (Soret et al., 2017) reached peak T conditions at 96–95 Ma
384 (Fig. 4a), consistent with zircon U–Pb dates of 96.2–94.5 Ma for melt segregations in HTa
385 (Rioux et al., 2016). Therefore, both units appear to have been detached from the slab
386 synchronously with upper plate extension and associated spreading of the supra-subduction
387 zone ophiolite at 96.5–95.5 Ma.

388

389 In addition, the prograde record in titanite and monazite shows that amphibolite-facies
390 conditions of ~ 650 – 750°C were reached by (102–)100 Ma (Fig. 4a, 7c). These new data range

391 amongst the older set of dates published for the Semail metamorphic sole: the 104–103 Ma Lu–
392 Hf garnet dates for HTa (Guilmette et al., 2018) (Fig. 7a), and the 107–102 Ma U-Pb dates of
393 zircon cores found in a unit equivalent to HTb (Garber et al., 2020). In the latter study, the
394 zircon rims with low Th/U ratios (dated at 98.7–94.1 Ma) were reported to co-exist with Th-
395 rich monazite (dated at 97–89 Ma). The 107-102 Ma zircon cores with high Th/U ratios without
396 significant Eu anomalies may also have formed along the prograde P-T path prior to the
397 crystallization of monazite rather than representing detrital cores as interpreted.

398
399 Titanite U-Pb dates ranging from 100 to 96 Ma indicate that the Lu–Hf dates from the
400 chemically zoned garnet (Guilmette et al., 2018) may actually reflect continuous garnet growth
401 from > 104 Ma to 96 Ma. Significant isobaric heating is recorded in HTa during peak burial
402 garnet growth (>100°C; Fig. 7c), as reflected by phase equilibrium modeling (Soret et al.,
403 2017), increasing average Zr content (from 235 to 580 ppm; Fig. 4a), and steepening of the
404 REE pattern in titanite included in garnet. Consequently, HTa was not yet detached from the
405 slab at 104–103 Ma and was therefore not accreted to the upper plate at supra-solidus conditions
406 for 8 Myr, contrary to previous claims (Guilmette et al., 2018). Zoning in garnet (Garber et al.,
407 2020; Guilmette et al., 2018; Soret et al., 2017) and titanite (Fig. SM4) would also less likely
408 be preserved over such long periods (Cherniak, 2006). Furthermore, the Zr decrease in titanite
409 grains from the partly retrogressed matrix (SE14-31, G5) confirms the detachment of the HTa
410 metamorphic sole unit from the slab at 95 Ma, immediately after reaching peak P-T conditions.

411
412 This study quantifies the timescales for the transition from subduction nucleation to mature
413 subduction (“slabitization” stage; Agard et al., 2020; Fig. 7a) with emphasis on the evolution
414 of the force balance during convergence (Gurnis et al., 2004; Lallemand and Arcay, 2021;
415 Shuck et al., 2022).

416

417 The extensive U–Th/Pb dataset presented here confirms that subduction initiation predated
418 upper plate extension at 96–95 Ma. It follows that subduction initiation was tectonically
419 induced, probably by a major plate reorganization and plate acceleration event at 115–105 Ma
420 (Agard et al., 2007; Matthews et al., 2015; Rodriguez et al., 2021; van Hinsbergen et al., 2021).
421 From at least 102–100 Ma onwards (and probably as early 107–104 Ma) to 97–96 Ma, the future
422 HTa and HTb soles were slowly subducted until peak burial (at 1.0 and 0.7–0.8 GPa,
423 respectively) as shown by the warm excursion of the P–T path (Fig. 8c) and still attached to the
424 slab. Petrochronology, therefore, reveals that subduction initiated at a slow rate $< \text{cm/yr}$,
425 considering a 30° subduction dip angle, for 5–10 Myr (from t_0 to t_1 ; Fig. 8a, b). We refer to
426 this period of slow subduction progress (i.e. rate $< \text{cm/yr}$; see also Anczkiewicz et al. (2004)
427 for the Franciscan example) as “stagnation” (Fig. 8a, 9). The lack of subduction to greater
428 depths is supported by the absence of pre-96 Ma subduction-related magmatism (Belgrano and
429 Diamond, 2019; Rioux et al., 2021b, 2016).

430

431 This evolution can be explained by the combination of two resisting forces: the initial elastic
432 flexure of the downgoing plate and the onset of viscous coupling at depth accompanying
433 induced subduction initiation (Fig. 8c, 9). The nascent slab first encountered the serpentinized
434 shallow mantle (at t_0 ; Fig. 8a–c) followed by progressively drier and mechanically stronger
435 mantle with increasing depth and temperature, which acted as a buttress (at t_1 , Fig. 8b; Agard
436 et al., 2016): when the slab tip eventually reached a viscosity similar to that of the overlying
437 metasomatized peridotites it became fully coupled with the warm mantle wedge (Agard et al.,
438 2016). We relate garnet nucleation and growth in the metamorphic sole at ≥ 104 –98 Ma (at
439 $T \sim 650$ – 750°C , ~ 30 km depth; Fig. 7c, 8c) to this stage of thermal and mechanical viscous
440 coupling between the nascent plates (stagnation; Fig. 8a–c, 9).

441

442 Extensive slab dehydration along the coupled interface during stagnation (Fig. 8b, 9), as
443 predicted by phase equilibrium modeling (Fig. 4), was critical for subduction progress. It
444 promoted (i) viscous decoupling, through progressive hydration and viscosity decrease of the
445 mantle wedge (Agard et al., 2016; Soret et al., 2019a) and (ii) slab pull, through slab
446 densification (Hacker et al., 2003) induced by garnet-forming dehydration reactions (Fig. 8b).
447 This combined thermal and mechanical evolution overcame mantle resistance, ending
448 stagnation and triggered the fast “unlocking” of the nascent plate interface (~1 Ma; t₂; Fig. 8a-
449 c). Subduction unlocking coincided with a significant increase in the convergence rate ($\geq 2\text{--}5$
450 cm/yr; Fig. 9; Boudier et al., 1988; Duretz et al., 2016) and induced local, fast mantle upwelling
451 and upper plate extension leading to spreading of a new lithosphere in ≤ 2 Myr (Rioux et al.,
452 2021b, 2016) above the nascent subduction zone (Belgrano and Diamond, 2019; Godard et al.,
453 2006; Haase et al., 2016). Associated heating (at t₂; Fig. 7c, 8c) promoted slab melting (Fig.
454 8b), which explains the presence of *in situ* melts (Fig. 2h) and felsic intrusions in the mantle
455 wedge (Haase et al., 2015; Rioux et al., 2021b, 2021a, 2016), both dated between 96 and 90
456 Ma. Mantle upwelling also facilitated the rapid exhumation of the metamorphic sole from 96–
457 95 Ma onwards (t₂; Figs. 8b, 9). Subduction rapidly evolved into a cold, self-sustained regime
458 after the detachment of the last unit of metamorphic sole (LT sole; ~92 Ma; Fig. 8c) to reach
459 thermal regimes similar to those modeled by van Keken et al. (2011) – characterized by evolved
460 arc magmatism dated at 87 Ma (Bonnet et al., 2020).

461

462 This evolution reconciles otherwise conflicting geochronology and subduction initiation
463 models in terms of mechanics, kinematics and thermal evolution. In particular, there is
464 considerable agreement with dynamic models of present-day incipient subduction systems, in
465 which the transition from forced to self-sustaining states is a 5–15 Myr-long continuous but

466 non-linear process requiring a critical amount of convergence and fluid released to overcome
467 the initially large resisting forces (Gurnis et al., 2004; Lallemand and Arcay, 2021; Mao et al.,
468 2017; Shuck et al., 2022). Although the critical convergence is a function of the age of the two
469 plates, its average value of ~100 km in the dynamic models is consistent with that calculated
470 for the Semail subduction system at the time metamorphic soles reached peak-P conditions
471 (considering a 30° dip angle). Moreover, the interpretation that supra-subduction zone
472 ophiolites form above a cooling plate interface during advanced stages of the subduction history
473 (“unlocking” stage; Fig. 9) explains the occurrence of blueschist-facies metamorphism (~1.6
474 GPa and 590°C at 51–48 Ma; Tamblyn et al., 2019) following the earliest stages of supra-
475 subduction zone magmatism (52–48 Ma; Ishizuka et al., 2011), as reported in the Izu–Bonin–
476 Mariana forearc (i.e., the most-studied example of *in situ* subduction initiation).

477
478 Finally, P-T-t paths also show that the initial subduction thermal regime was colder than
479 previously thought, at ~15–20°C/km (versus 25–30°C/km; t₀, t₁; Fig. 8c). Downward
480 migration of viscous coupling and serpentinite stability, at approximately constant temperature,
481 shifted subduction gradients towards colder values (t₂; Fig. 8c). Our model thereby bridges the
482 gap between subduction initiation and mature subduction and explains the transition from
483 decoupling to full coupling (≥ 80 km) observed in active subduction zones across the same
484 range of slab surface temperatures (t₂; Fig. 8c, 9; Syracuse et al., 2010).

485

486 **5. Conclusions**

487 This petrochronology study of the Semail ophiolitic complex (Oman-UAE) provides
488 information over the dynamics (mechanics, kinematics and thermal evolution) and duration of
489 subduction initiation. It confirms that subduction of the Arabian Plate, which ultimately led to
490 the obduction of the Semail ophiolite, was tectonically-induced by far-field forces prior to 100

491 Ma. It also shows that subduction started at a slow pace for ~ 10 Myr, with a closure rate < 1
492 cm/yr in increasingly warmer conditions, between ≥ 104 –100 and 95 Ma (“stagnation” stage)
493 and triggered incipient mantle convection and fast lithosphere formation at 96–95 Ma
494 (“unlocking stage”), eventually leading to a faster (2–5 cm/yr) and colder self-sustained regime.
495

496 The rock record of the Semail ophiolite as read through high-resolution geochronology shows
497 how thermo-mechanical changes of the plate interface allow subduction to proceed. Integrating
498 these results into active subduction zone datasets reconciles conflicting models of subduction
499 initiation and provides timescales for the evolution from subduction nucleation to (cold) mature
500 subduction. It also provides a coherent scheme to understand inception of self-sustained
501 subduction on Earth with modern-day plate tectonics.

502

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511

512 **AUTHOR CONTRIBUTIONS**

513 M.S. conceived the project and completed the petrological study. M.S and G.B. conducted and
514 processed the U–Pb geochronological analyses, and are the principal authors of the manuscript.
515 M.S, P.A. and B.D. carried out the field work. P.A. and K.L. co-supervised the project. J.C. and

516 A.K-C. helped in the acquisition of geochronological analyses. M.B and N.R. helped in the
517 acquisition of the petrological data. All authors contributed to discussion of the results and to
518 the manuscript.

519

520 **COMPETING FINANCIAL INTERESTS STATEMENT**

521 The authors declare no competing interests.

522

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766 **FIGURE CAPTIONS**

767 **Figure 1.** a) Conceptual model for the metamorphic sole and ophiolite formation during
768 subduction initiation (after Agard et al., 2016, and Guilmette et al., 2018). (b) Geodynamic
769 setting representing the evolution of oceanic subductions from birth to mature (cold) stage
770 (“slabification stage”; after Agard et al., 2020). HTa, HTb and LT refer to the three main slivers
771 of metamorphic sole accreted to the upper plate during subduction (Soret et al., 2017).

772

773 **Figure 2.** (a) Simplified geological map of the Semail ophiolite, highlighting the outcrops of
774 metamorphic sole (modified after Nicolas et al., 2000); (b) Schematic log of the Semail
775 ophiolite showing the structural position of the specimens from Khubakhib (K) and Sumeini
776 (S) (modified after Soret et al., 2017). See Table 1 for details.

777

778 **Figure 3.** Petrography of the analyzed specimens. (a-f) Photographs of titanite in the mafic HTa
779 and HTb units of the Semail metamorphic sole; (g-k) Back-scattered images of a peritectic
780 garnet with nanogranitoid inclusions (g-i), and monazite (j,k) in a metasedimentary layer at the
781 top of HTb. Mineral abbreviations are after Whitney and Evans (2010).

782

783 **Figure 4.** Phase equilibria modelling for (a-b) a garnet-bearing (mafic) amphibolite (SE16-26,
784 Sumeini) and (b) a metasedimentary (felsic) rock (SE13-123L, Khubakhib) from HTb. (a) T–
785 X(H₂O) calculations for the mafic specimen at 0.75 GPa with water content comprised between
786 0.5 and 2.0 wt%. Inset shows the P–T estimates calculated from the amphibole-plagioclase
787 thermobarometer by Holland and Blundy (1994) and Molina et al. (2015). Arrows highlight the
788 inferred P–T path between garnet core and rim growth.

789

790 **Figure 5.** (a) Correlation between the titanite age populations and the garnet zoning in HTa.

791 The grey envelop and the red arrow outline the prograde zoning of garnet determined from
792 different specimens of HTa (see Soret et al. (2017) for details). (b-r) Rare Earth element spider
793 diagrams of titanite colored by Zr content (ppm) for each laser ablation spot. $^{206}\text{Pb}^*/^{238}\text{U}$
794 weighted mean dates are given ± 2 SE (see Fig. SM2 for details). Average Zr contents (ppm)
795 are given ± 1 SE.

796

797 **Figure 6.** (a-b) $^{208}\text{Pb}^*/^{232}\text{Th}$ versus $^{206}\text{Pb}^*/^{238}\text{U}$ concordia plot of age data obtained from
798 monazite in specimen SE13-123L (Khubakhib). All isotopic ratios were corrected for common
799 Pb using the method detailed in the Method section. Two populations (G1 and G2) were
800 distinguished based on their composition. (c-d) Monazite Rare Earth element spider diagrams
801 for G1 and G2 (respectively) versus the Pb*/Th ages (± 2 SE).

802

803 **Figure 7.** Petrochronological results. (a) Zr content (ppm) and associated calculated
804 temperatures (thermometer of Hayden et al., 2008) vs. $^{206}\text{Pb}^*/^{238}\text{U}$ dates (± 2 SE) of titanite
805 from HTa (in red) and HTb (in blue). (b) Y content (ppm) (± 1 SE) vs. $^{208}\text{Pb}^*/^{232}\text{Th}$ dates
806 weighted mean ages for monazite grains comprising more than 6 spot analyses (see Table 1 for
807 details). (a-b) red and blue arrows highlight the prograde (P–)T–t paths of both metamorphic
808 sole units during subduction infancy. Grey arrow outlines the retrograde history. The results
809 are consistent with the thermodynamic modelling conducted in this study and by Soret et al.
810 (2017). Major geodynamical events are outlined in grey for comparison. GPRE: global plate
811 reorganization event. (c) Summary of P–T–t paths for each HTa and HTb. Ref.(1) : this study,
812 (2) Guilmette et al. (2018), (3) Rioux et al. (2016), (4) Garber et al. (2020), (5) Hacker et al.
813 1996, (6) Matthews et al. (2015), (7) Rioux et al. (2021b).

814

815 **Figure 8.** Thermo-mechanical model for the early subduction dynamics. (a) Timeline

816 highlighting the major stages (t0, t1, t2) of the oceanic subduction, from birth to death. Three-
817 stage mechanical (b) and thermal (c) evolution reconciling bridging the gap between subduction
818 initiation dynamics and that of mature subduction zones. Markers 'a (HTa), b (HTb), c, d' are
819 used to trace the slab evolution during the subduction history. Symbol (*) refers to ref. Agard
820 et al. (2020).

821

822 **Figure 9.** Geodynamic setting representing the timescales of the Neotethys oceanic subduction
823 next to Arabia. The evolution from subduction nucleation to (cold) mature subduction
824 ("slabification"; Agard et al., 2020) is first controlled by a stage of "stagnation" (104–97 Ma)
825 followed by a stage of "unlocking" (96–92 Ma).

826

827 **Table 1.** U–Th/Pb dates and compositions of titanite and monazite from the Semail
828 metamorphic sole.

829

830 **FIGURE AND TABLE CAPTIONS FOR SUPPLEMENTARY INFORMATION**

831 **Document SM1.** Detailed description of all the methods used in this study.

832 **Figure SM1.** Weighted mean ^{207}Pb -corrected $^{206}\text{Pb}/^{238}\text{U}$ dates of titanite. Weighted mean
833 values are outlined by the black horizontal line, with the 95% confidence interval shown as a
834 grey band. Spot analyses are plotted ± 2 SE, in red for HTa and in blue for HTb. Outliers
835 rejected from the weighted mean calculations are shown as white bars.

836 **Fig. SM2.** BSE images of the dated titanite grains highlighting their textural positions and the
837 laser ablation spot locations.

838 **Figure SM3.** Compositional map of Yttrium (number of counts) of monazite in specimen
839 SE13-123L with LASS spot locations and related $^{208}\text{Pb}^*/^{232}\text{Th}$ dates.

840 **Figure SM4.** Chemical map showing the Y and Zr zoning in a large titanite included in garnet

841 (SE13-123A).

842 **Table SM1.** Titanite U–Pb and Trace element analysis.

843 **Table SM2.** Monazite U–Th/Pb and Trace element analysis.

844 **Table SM3.** Bulk rock compositions (%wt) used for phase equilibrium modeling.

845

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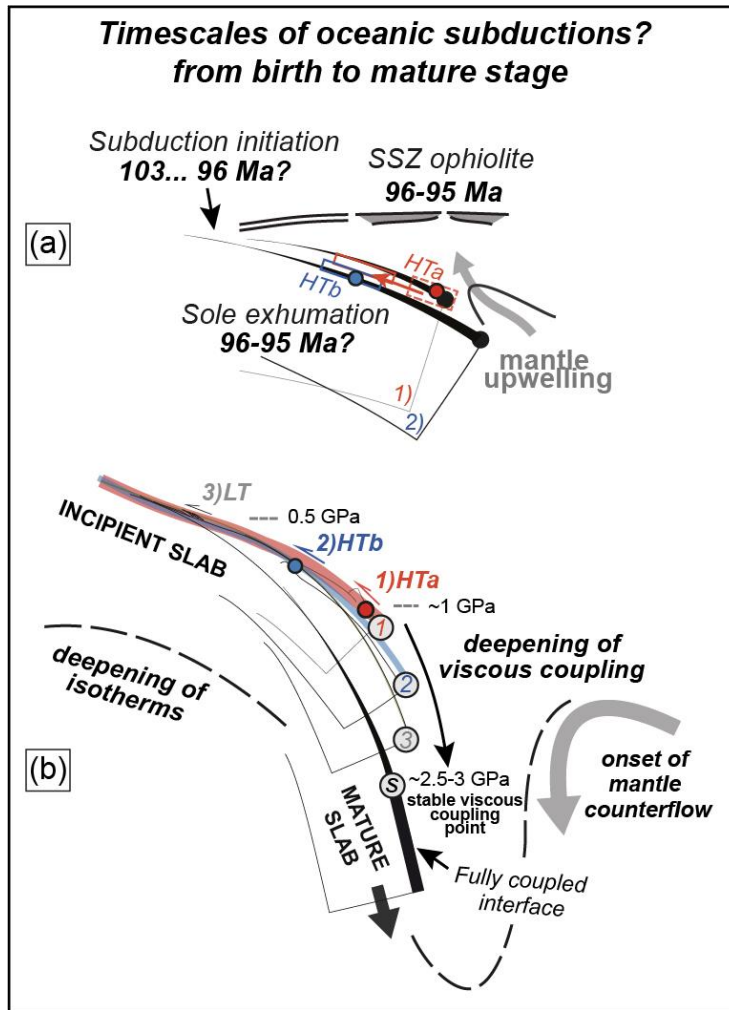


Figure 1

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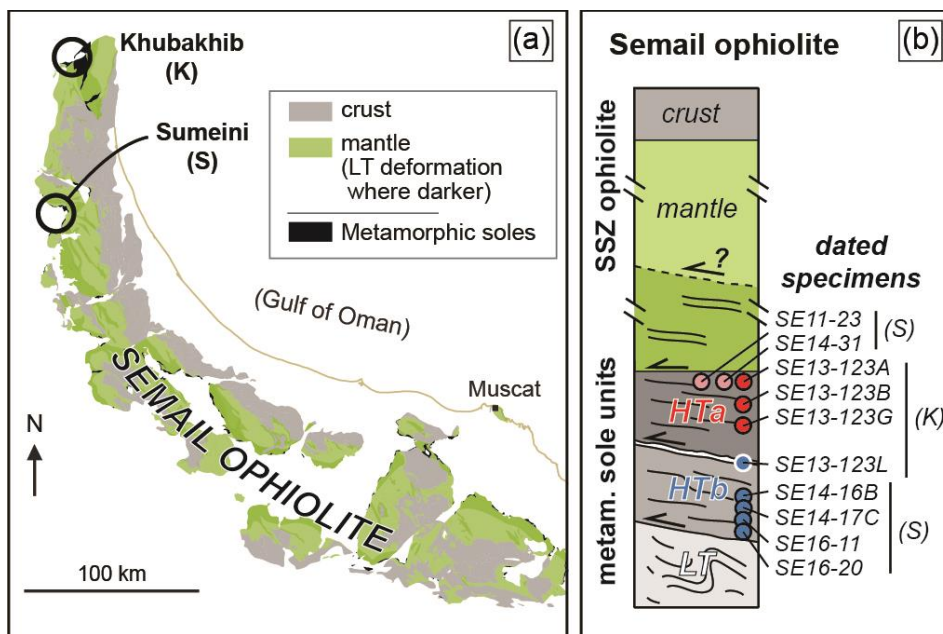
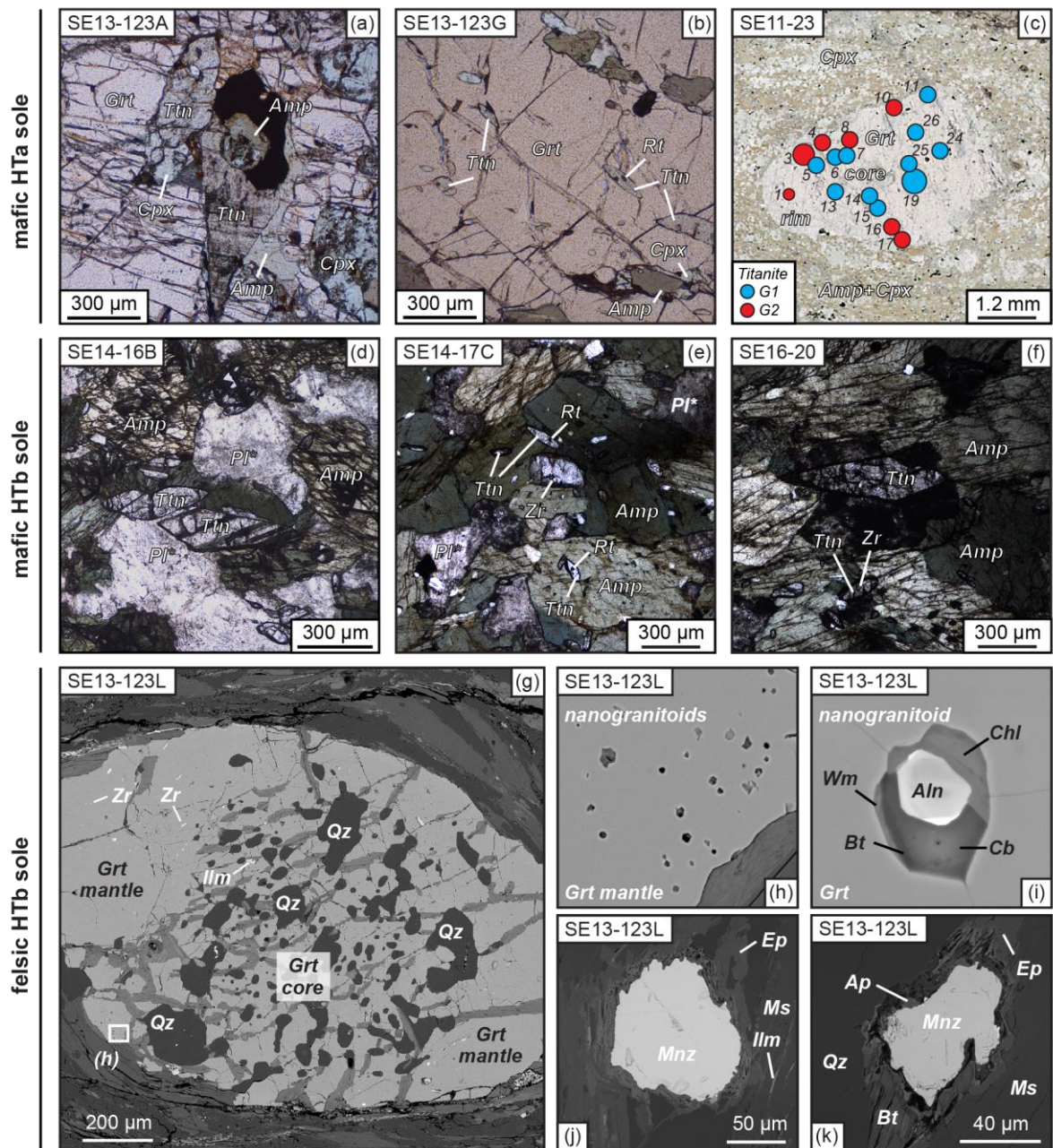


Figure 2

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Figure 3

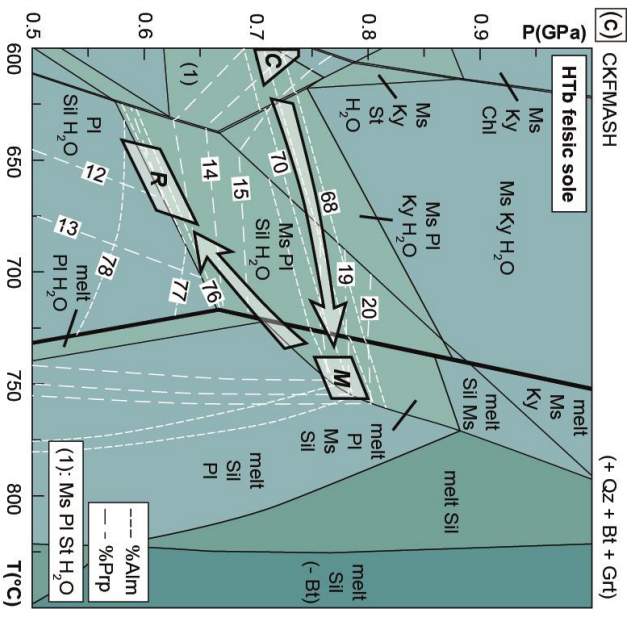
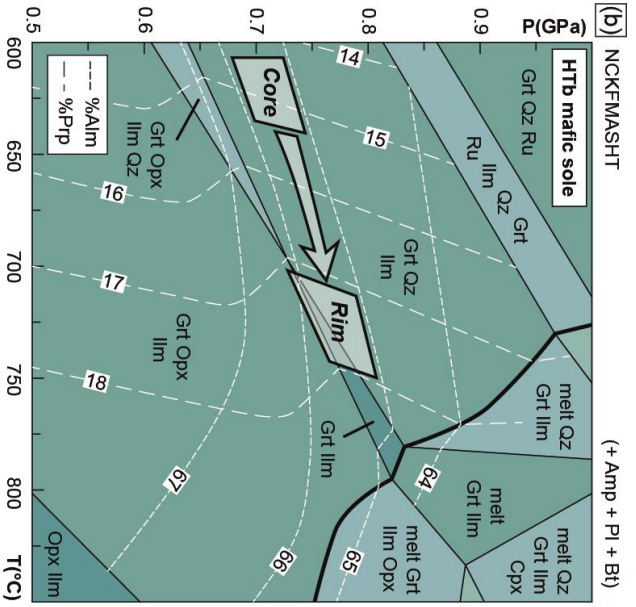
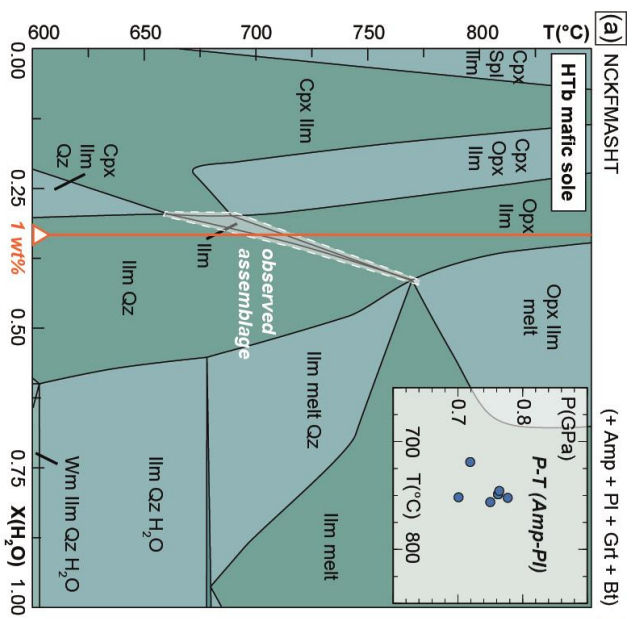


Figure 4

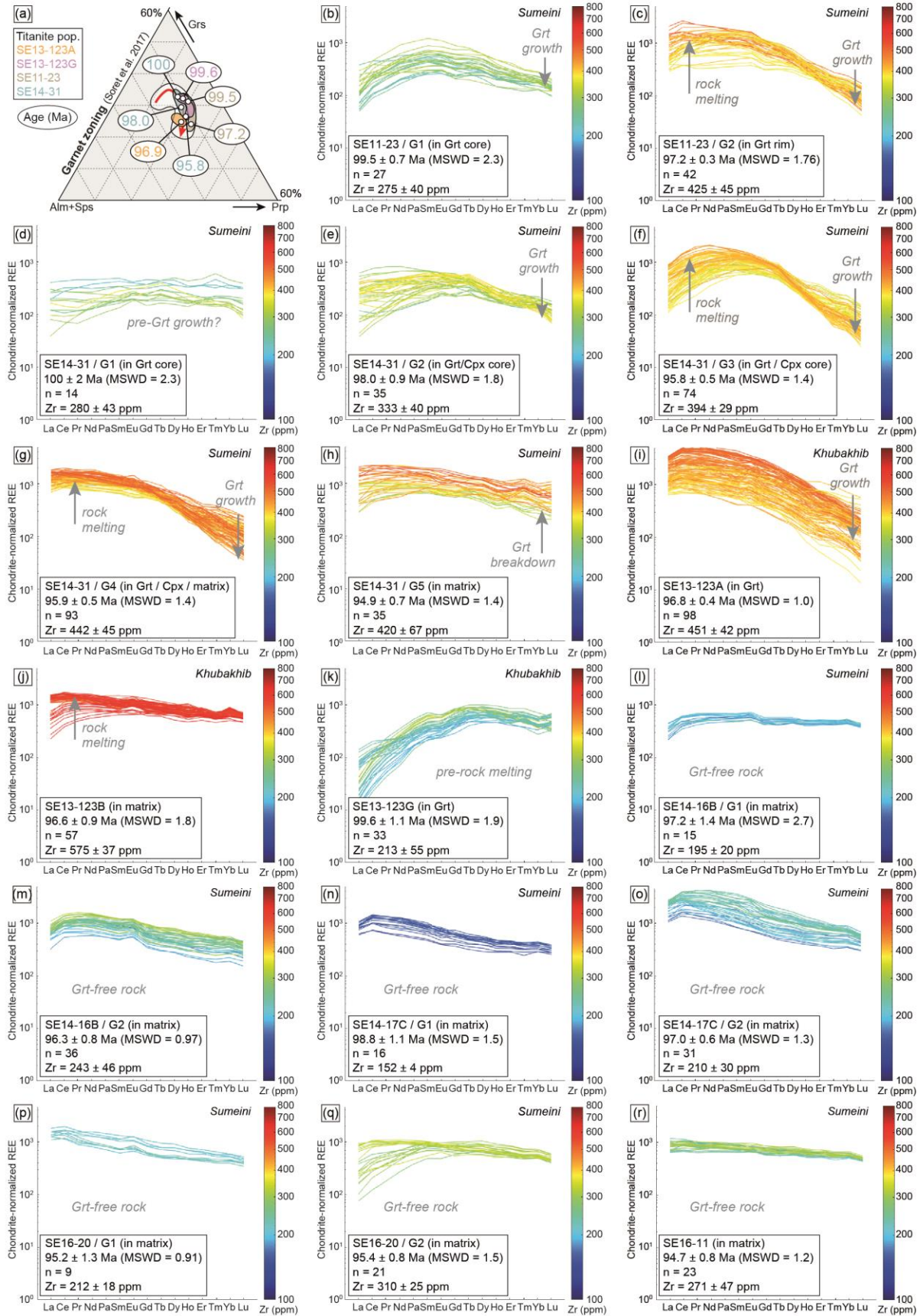


Figure 5

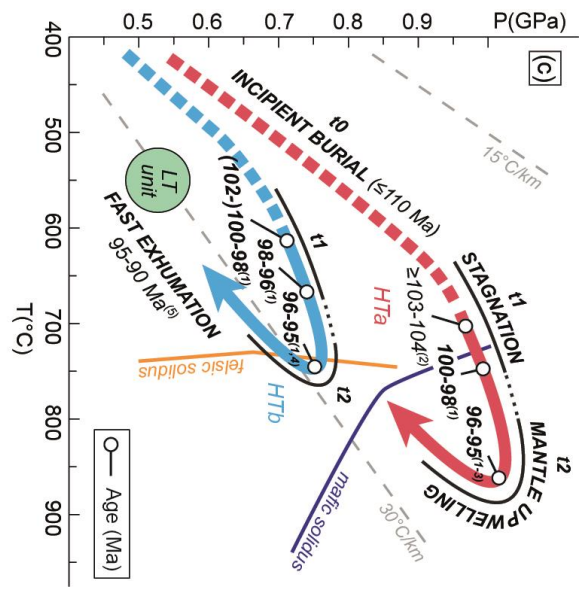
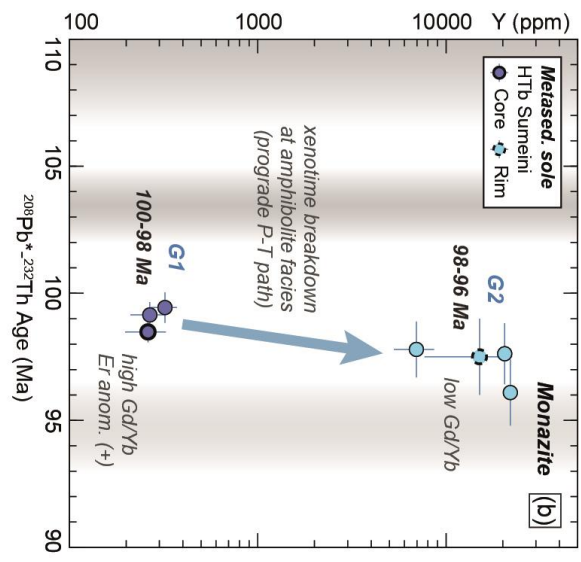
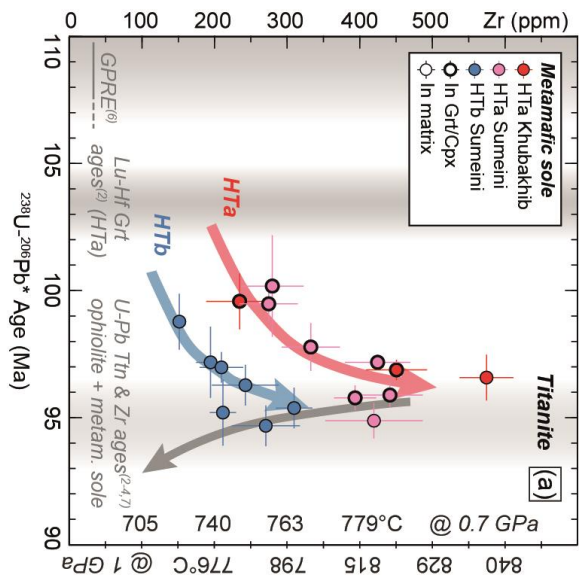


Figure 7

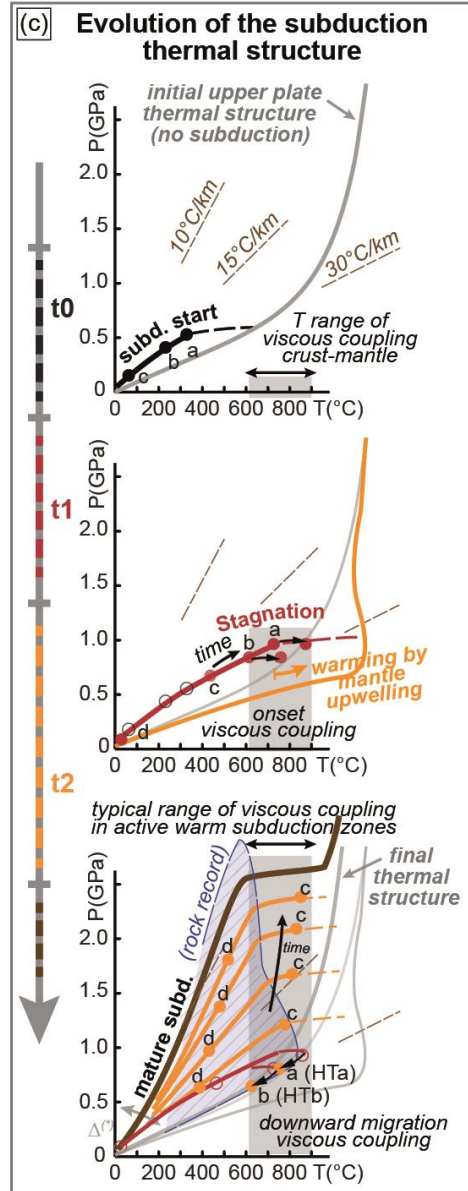
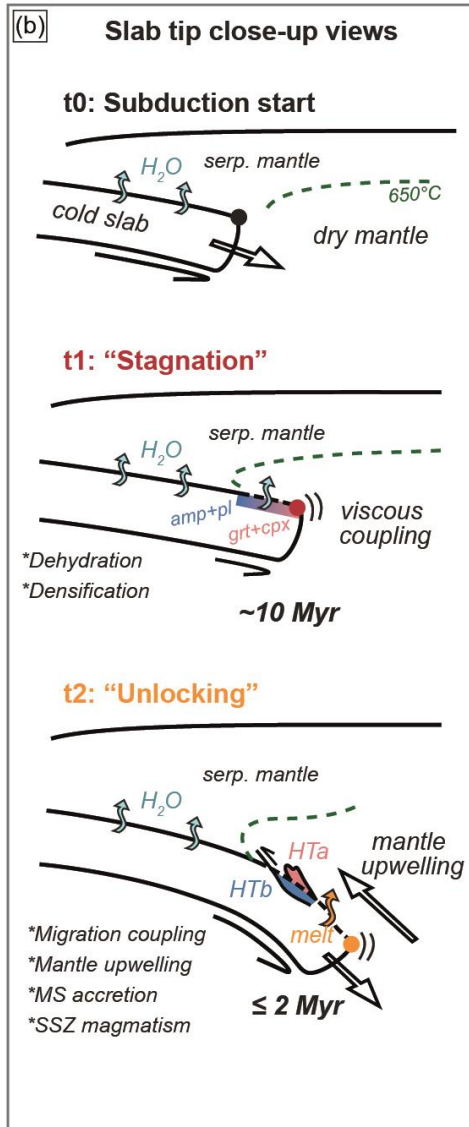
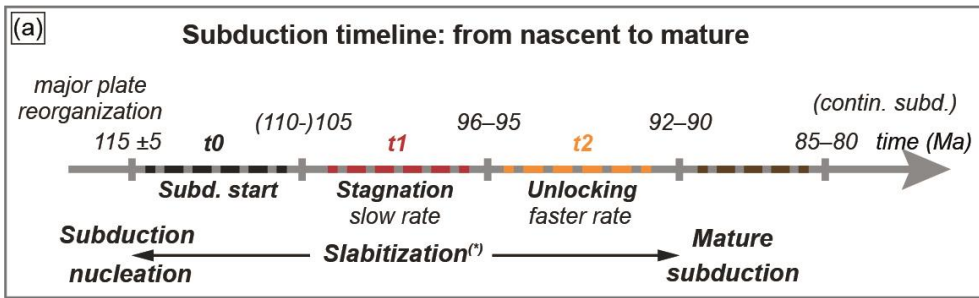


Figure 8

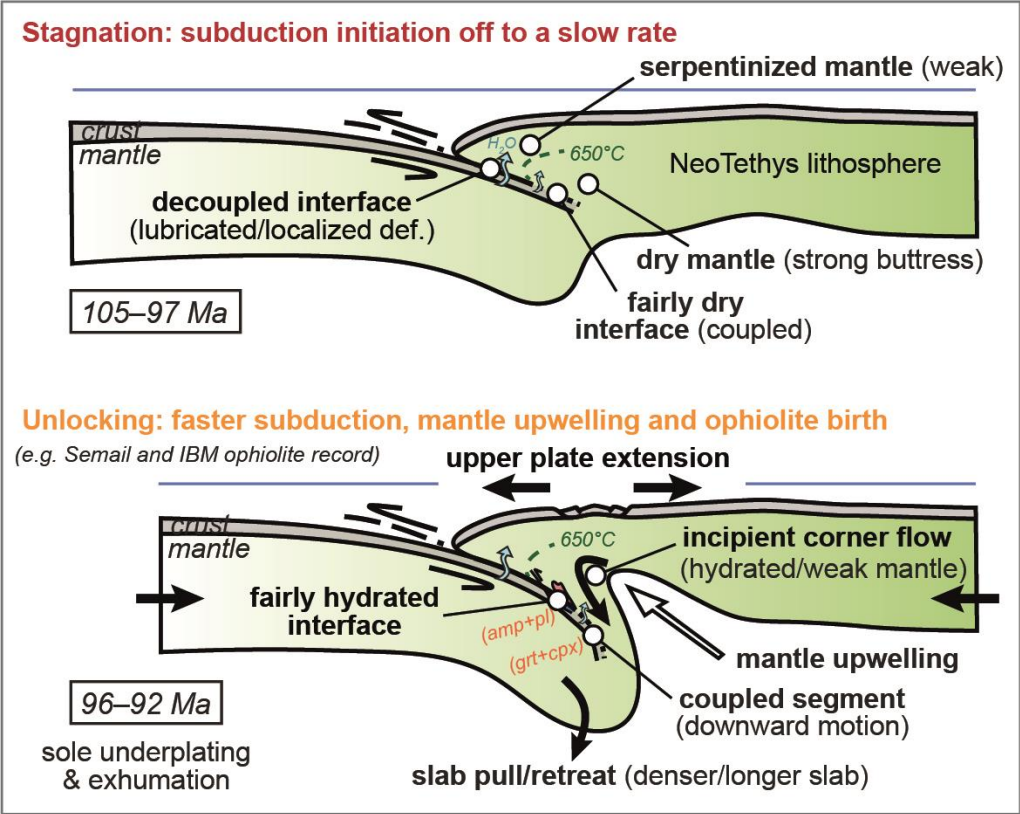


Figure 9

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Table 1
U–Th/Pb dates and compositions of titanite and monazite from the Semail metamorphic sole.

Dated mineral	System	Sample	Unit	Locality	Textural position	# ⁽¹⁾	Weighted mean date ⁽²⁾	W. mean MSWD	Average Zr (ppm)	Average Y (ppm)	Average Gd/Yb
Titanite	U-Pb	SEI3-123A	HTa	Khubakhib	in Grt	98	96.9 ± 0.4	1.0	451 ± 42		
		SEI3-123B	HTa	Khubakhib	in matrix	57	96.6 ± 0.9	1.8	575 ± 37		
		SEI3-123C	HTa	Khubakhib	in Grt	33	99.6 ± 1.1	1.9	235 ± 46		
		SEI4-31	HTa	Sumeini	in Grt (G1)	14	100 ± 2	2.3	280 ± 43		
					in Grt and Cpx (G2)	35	98.0 ± 0.9	1.8	333 ± 40		
					in Grt and Cpx (G3)	74	95.8 ± 0.5	1.4	394 ± 29		
					in Grt, Cpx and matrix (G4)	93	95.9 ± 0.5	1.5	425 ± 45		
					in matrix (G5)	35	94.9 ± 0.7	1.4	420 ± 67		
					in Grt core (G1)	27	99.5 ± 0.7	2.3	275 ± 40		
					in Grt rim (G2)	42	97.2 ± 0.3	1.8	425 ± 45		
Titanite	U-Pb	SEI4-16B	HTb	Sumeini	in matrix (G1)	15	97.2 ± 1.4	2.7	195 ± 20		
					in matrix (G2)	36	96.3 ± 0.8	1.0	243 ± 46		
		SEI4-17C	HTb	Sumeini	in matrix (G1)	16	98.8 ± 1.1	1.5	152 ± 4		
					in matrix (G2)	31	97.0 ± 0.6	1.3	210 ± 30		
		SEI6-11	HTb	Sumeini	in matrix	23	94.7 ± 0.8	1.2	271 ± 47		
					in matrix (G1)	9	95.2 ± 1.3	0.9	212 ± 18		
Monazite	Th-Pb	SEI6-20	HTb	Sumeini	in matrix (G2)	21	95.4 ± 0.8	1.5	310 ± 25		
		SEI3-123L1	HTb	Sumeini	in matrix Z9 (G2)	7	97.6 ± 1.2	1.0	20417 ± 1420	23 ± 2	
		SEI3-123L2	HTb	Sumeini	in matrix Z1 (G1)	27	99.2 ± 0.5	1.4	265 ± 55	2180 ± 360	
		SEI3-123L2	HTb	Sumeini	in matrix Z3 (G2)	7	96.6 ± 1.3	0.6	21867 ± 1686	27 ± 4	
		SEI3-123L2	HTb	Sumeini	in matrix Z8 (G2)	9	97.8 ± 1.1	1.1	6923 ± 1661	38 ± 4	
		SEI3-123L2	HTb	Sumeini	in matrix Z9 (G1)	23	99.5 ± 0.6	1.3	320 ± 50	2360 ± 520	
		SEI3-123L2	HTb	Sumeini	in matrix Z10, grain core (G1)	48	98.5 ± 0.4	1.0	260 ± 60	2670 ± 650	
		SEI3-123L2	HTb	Sumeini	in matrix Z10, grain rim (G2)	6	97.5 ± 1.5	0.1	15000 ± 7340	41 ± 18	

(1) Number of spot analyses; (2) Weighted mean ± 2 SE (Ma); MSWD: Mean Squared Weighted Deviation; Trace element composition ± 1 SE.