

Timescales of subduction initiation and evolution of subduction thermal regimes

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1	Timescales of subduction initiation and evolution of subduction
2	thermal regimes

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37 Highlights

- Titanite and monazite in metamorphic soles record subduction initiation dynamics
- 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 which subduction initiates remain enigmatic and controversial. Here, we reappraise the timing 46 47 of metamorphism of the rock units first detached from the leading edge of the downgoing slab during initiation of the Neotethys subduction, now preserved in the metamorphic sole of the 48 49 Semail ophiolite (Oman-United Arab Emirates). Using petrochronology and phase equilibrium modeling, we demonstrate that subduction initiated prior to 102-100 Ma at a slow rate (< 1 50 cm/yr). Subduction stagnated at relatively warm conditions $(15-20^{\circ}C/km)$ for > 5 Myr before 51 52 evolving into a faster ($\geq 2-5$ cm/yr) and colder ($\sim 7^{\circ}$ 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 analogue and numerical subduction initiation models, shedding light on the thermal, mechanical 56 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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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; Crameri et al., 2020; Gurnis et al., 2004; Holt and Condit, 74 2021). Direct geological records of subduction initiation are limited to metamorphic soles, which correspond to thin sheets of deformed and metamorphosed oceanic crust. Metamorphic 75 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 remnants of the tip of the downgoing slab metamorphosed and deformed at high temperature 78 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).

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 (> cm/yr) along 105 a pre-existing lithospheric-scale weak zone.

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⁴⁰Ar/³⁹Ar and U-Pb geochronology of metamorphic soles and associated ophiolites (Garber et 107 108 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 (~850°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~650°C) confirms that prograde metamorphism started \geq 98.7 ± 1.7 Ma (Garber et al., 2020). 117 Dates from the cores of the same zircon grains $(107-102 \pm 2 \text{ Ma})$, though coeval within

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

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

128

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

135

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

The Semail metamorphic sole comprises a dominantly mafic, high-temperature (HT) unit thrust onto a mainly metasedimentary low-temperature (LT) unit (Fig. 2a, b; Gnos, 1998). The HT sole itself consists of two tectonic slices juxtaposed sequentially, HTa then HTb (Soret et al., 2017). The latter is capped by a meter-thick garnet-rich metasedimentary layer. Specimens were 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**

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

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Amphibolite-facies metabasalts of HTb mainly consist of amphibole and plagioclase with minor biotite and either clinopyroxene or chemically zoned garnet (Garber et al., 2020; Soret et al., 2019a, 2017). Titanite forms euhedral to subhedral grains closely associated with both amphibole and plagioclase, either as inclusions or along boundaries (Fig. 3e-f). Rutile and zircon form small inclusions (1–30 μ m) in titanite and amphibole (Fig. 3e). In the presence of 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 µm) and typically inclusion-free, save for minute 166 monazite and zircon $(1-10 \ \mu m)$. Large monazite $(50-100 \ \mu m;$ Fig. 3j, k) occurs in the rock 167 matrix where it is rimmed by a syn-kinematic assemblage of epidote, allanite and apatite. 168

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

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177 **2.2. Existing geochronology**

Zircon U-Pb dating of melt segregations in HTa indicates melt crystallization and incipient 178 cooling by 96.2–94.5 Ma (Guilmette et al., 2018; Rioux et al., 2016). Hornblende ⁴⁰Ar/³⁹Ar 179 180 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 ${}^{206}Pb/{}^{238}U$ age of 95.6 \pm 0.3 Ma, which was 183 interpreted as a cooling age below 600-650°C (Guilmette et al., 2018). Finally, titanite and monazite U-Pb dating combined with garnet Lu-Hf dating in a lower grade unit (peak 184 185 T~650°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).

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**

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

Peak P–T conditions inferred for the Grt + Amp + Pl + Bt + Ilm \pm Qz assemblage of the garnetbearing amphibolite (SE16-26, Sumeini; Fig. 4a, b) lie at sub-solidus conditions of 725 \pm 25°C and 0.77 \pm 0.03 GPa. In addition to the calculated garnet rim composition (Alm₆₅₋₆₆Prp₁₇₋₁₈Grs₁₇), calculated modes (Grt₁₀Amp₃₅Pb₈Bt₁₀Ilm_{2.5}) and plagioclase composition (An₄₁) are in close agreement with those observed in the specimen (Soret et al., 2017). Garnet core isopleths converge at 625 \pm 20°C and 0.70 \pm 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 $750 \pm 10^{\circ}$ 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 for the peritectic garnet mantle (Alm₆₉₋₇₁Prp₁₈₋₁₉Grs₁₂₋₁₃). The composition of the quartz-rich 229 230 garnet core (Alm₆₇₋₆₈Prp₁₃₋₁₄Grs₁₈₋₁₉) indicates a prograde, near-isobaric evolution from $610 \pm$ 231 10°C and 0.72 \pm 0.2 GPa. The thin inclusion-free garnet rim (Alm₇₆₋₇₇Prp₁₂Grs₁₁) further 232 indicates sub-solidus re-equilibration during retrograde P-T conditions at $660 \pm 20^{\circ}$ 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).

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238

8 **3.2.** *In situ* titanite petrochronology

In situ U–Th/Pb and trace-element analyses of titanite and monazite (see section 3.3) were carried out on five HTa specimens and five HTb specimens (Fig. SM2, SM3) using the LASS facility at the University of California, Santa Barbara (USA). Data reduction for U–Th/Pb, including corrections for baseline, instrumental drift, mass bias, down-hole fractionation were carried out using Iolite v. 2.5 (Paton et al., 2011), and error correlations were recalculated after Schmitz and Schoene (2007). Methods used in this study follow those outlined by Kylander-Clark et al. (2013), with modifications presented by McKinney et al. (2015).

246

Rutile and zircon, though present in all specimens, proved too small ($\leq 20 \,\mu$ m) to be analyzed. Different titanite age populations within each specimen in both HTa and HTb units were distinguished based on their textural position and Rare Earth elements (REE) contents (Fig. 5). Titanite ages were calculated for all titanite populations. They correspond to ²⁰⁷Pb-corrected (following Stacey and Kramers, 1975) ²⁰⁶Pb/²³⁸U weighted means, and they ranged between 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*

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

259

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

enrichment in Zr relative to G1, and yielded a date of 98.0 ± 0.9 Ma (n = 35). The third 266 267 population (G3) also comprises inclusions in the core of garnet and clinopyroxene, but has a distinct REE pattern marked by an enrichment in LREE, a larger HREE depletion, and higher 268 269 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 in LREE with a similar HREE depletion, and the highest Zr concentrations of any of the titanite 272 273 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 \pm 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

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

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. 51-r). Different 294 age populations within each specimen were distinguished based on variations in Zr and REE 295 concentrations. In SE14-16B (Fig. 51-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 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. 297 298 5p-q), G1 returned a date of 95.2 \pm 1.3 Ma (n = 9), while G2 gave a date of 95.4 \pm 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 (\pm 25) ppm, with the highest concentration in the 96-95 Ma old population 304 (Fig. 51-r; Tab. 1).

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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 collected across 18 grains yielding most ²⁰⁸Pb*/²³²Th dates between 102 Ma and 96 Ma (Fig. 310 311 6a-b).

312

313 Two monazite age populations are recognized based on their REE patterns (Fig. 6c-d). The first population (G1) contains high Gd/Yb ratios (2180-2670) and low Y content (260-320 ppm). 314 315 Three grains (including the core of a zoned grain) were large enough for multiple analyses and 316 yielded weighted mean 208 Pb*/ 232 Th dates of 99.2 ± 0.5 Ma (n = 27, SE13-123L2, Z1), 99.5 ± 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

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

331

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

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In HTa, the negative slope in REE of titanite inclusions in the core of chemically zoned garnet and clinopyroxene porphyroclasts (Fig. 5) confirms they crystallized in equilibrium with garnet. The Zr-poor titanite population (200–300 ppm) yields temperature estimates (750–800°C, at P 341 $\sim 0.7-1.0$ 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) are calculated from the Zr-rich titanite population (400-600 ppm). Because temperature 347 348 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

In HTb, the average Zr content in titanite is systematically lower than in HTa at a given age, 357 358 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 yields a maximum temperature of 740-760°C (at P ~0.7 GPa), in agreement with peak 360 361 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 ~725°C (at P ~0.7 GPa), consistent with crystallization along the prograde path near peak 364 365 conditions and a lack of further re-equilibration of Zr.

367 Lead diffusivity in monazite is thought to be negligible up to temperatures of >800°C for grains 368 larger than 20 µm 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

The results of this study indicate that the leading edge of the nascent slab preserved in the HTa and HTb metamorphic sole units (Soret et al., 2017) reached peak T conditions at 96–95 Ma (Fig. 4a), consistent with zircon U–Pb dates of 96.2–94.5 Ma for melt segregations in HTa (Rioux et al., 2016). Therefore, both units appear to have been detached from the slab synchronously with upper plate extension and associated spreading of the supra-subduction 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 $\sim 650-750^{\circ}$ C were reached by (102–)100 Ma (Fig. 4a, 7c). These new data range amongst the older set of dates published for the Semail metamorphic sole: the 104–103 Ma Lu– Hf garnet dates for HTa (Guilmette et al., 2018) (Fig. 7a), and the 107–102 Ma U-Pb dates of zircon cores found in a unit equivalent to HTb (Garber et al., 2020). In the latter study, the zircon rims with low Th/U ratios (dated at 98.7–94.1 Ma) were reported to co-exist with Thrich monazite (dated at 97–89 Ma). The 107-102 Ma zircon cores with high Th/U ratios without significant Eu anomalies may also have formed along the prograde P-T path prior to the 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

This study quantifies the timescales for the transition from subduction nucleation to mature subduction ("slabitization" stage; Agard et al., 2020; Fig. 7a) with emphasis on the evolution of the force balance during convergence (Gurnis et al., 2004; Lallemand and Arcay, 2021; Shuck et al., 2022). 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 (Agard et al., 2007; Matthews et al., 2015; Rodriguez et al., 2021; van Hinsbergen et al., 2021). 420 421 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 < cm/yr, 425 considering a 30° subduction dip angle, for 5–10 Myr (from t0 to t1; Fig. 8a, b). We refer to 426 this period of slow subduction progress (i.e. rate < 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 magnatism (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 t0; Fig. 8a-c) followed by progressively drier and mechanically stronger 435 mantle with increasing depth and temperature, which acted as a buttress (at t1, 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 \geq 104–98 Ma (at T~650-750°C, ~30 km depth; Fig. 7c, 8c) to this stage of thermal and mechanical viscous 439 440 coupling between the nascent plates (stagnation; Fig. 8a-c, 9).

442 Extensive slab dehydration along the coupled interface during stagnation (Fig. 8b, 9), as predicted by phase equilibrium modeling (Fig. 4), was critical for subduction progress. It 443 444 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 ($\geq 2-5$ 450 cm/vr; 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).

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This evolution reconciles otherwise conflicting geochronology and subduction initiation models in terms of mechanics, kinematics and thermal evolution. In particular, there is considerable agreement with dynamic models of present-day incipient subduction systems, in 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 ophiolites form above a cooling plate interface during advanced stages of the subduction history 472 473 ("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

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

485

486 **5.** Conclusions

This petrochronology study of the Semail ophiolitic complex (Oman-UAE) provides information over the dynamics (mechanics, kinematics and thermal evolution) and duration of subduction initiation. It confirms that subduction of the Arabian Plate, which ultimately led to 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 $\geq 104-100$ 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

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

502

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511

512 AUTHOR CONTRIBUTIONS

M.S. conceived the project and completed the petrological study. M.S and G.B. conducted and
processed the U–Pb geochronological analyses, and are the principal authors of the manuscript.
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

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

772

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

777

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

782

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

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

796

Figure 6. (a-b) 208 Pb*/ 232 Th versus 206 Pb*/ 238 U concordia plot of age data obtained from monazite in specimen SE13-123L (Khubakhib). All isotopic ratios were corrected for common Pb using the method detailed in the Method section. Two populations (G1 and G2) were distinguished based on their composition. (c-d) Monazite Rare Earth element spider diagrams 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 temperatures (thermometer of Hayden et al., 2008) vs. ²⁰⁶Pb*/²³⁸U dates (± 2 SE) of titanite 804 from HTa (in red) and HTb (in blue). (b) Y content (ppm) (± 1 SE) vs. ²⁰⁸Pb*/²³²Th dates 805 806 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, (2) Guilmette et al. (2018), (3) Rioux et al. (2016), (4) Garber et al. (2020), (5) Hacker et al. 812 813 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

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

821

Figure 9. Geodynamic setting representing the timescales of the Neotethys oceanic subduction next to Arabia. The evolution from subduction nucleation to (cold) mature subduction ("slabitization"; Agard et al., 2020) is first controlled by a stage of "stagnation" (104–97 Ma) 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.

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

Fig. SM2. BSE images of the dated titanite grains highlighting their textural positions and thelaser ablation spot locations.

- Figure SM3. Compositional map of Yttrium (number of counts) of monazite in specimen
 SE13-123L with LASS spot locations and related ²⁰⁸Pb*/²³²Th 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.



Figure 1



















Figure 9

U-III/ED Udies and	composition	א טו תושוות: שוות ו		OIII UIG DEIIIMII I	incraintor brite 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	SE13-123A	HTa	Khubakhib	in Grt	86	96.9 ± 0.4	1.0	451 ± 42		
		SE13-123B	HTa	Khubakhib	in matrix	57	96.6 ± 0.9	1.8	575 ± 37		
		SE13-123G	HTa	Khubakhib	in Grt	33	99.6 ± 1.1	1.9	235 ± 46		
		SE14-31	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		
		SE11-23	HTa	Sumeini	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	SE14-16B	НТЪ	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		
		SE14-17C	НТЬ	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		
		SE16-11	НТЪ	Sumeini	in matrix	23	94.7 ± 0.8	1.2	271 ± 47		
		SE16-20	НТЬ	Sumeini	in matrix (G1)	9	95.2 ± 1.3	0.9	212 ± 18		
					in matrix (G2)	21	95.4 ± 0.8	1.5	310 ± 25		
Monazite	Th-Pb	SE13-123L1	НТЬ	Sumeini	in matrix Z9 (G2)	7	97.6 ± 1.2	1.0		20417 ± 1420	23 ± 2
		SE13-123L2	НТЪ	Sumeini	in matrix Z1 (G1)	27	99.2 ± 0.5	1.4		265 ± 55	2180 ± 360
		SE13-123L2	НТЪ	Sumeini	in matrix Z3 (G2)	7	96.6 ± 1.3	0.6		21867 ± 1686	27 ± 4
		SE13-123L2	НТЬ	Sumeini	in matrix Z8 (G2)	9	97.8 ± 1.1	11		6923 ± 1661	38 ± 4
		SE13-123L2	НТЬ	Sumeini	in matrix Z9 (G1)	23	99.5 ± 0.6	1.3		320 ± 50	2360 ± 520
		SE13-123L2	НТЬ	Sumeini	in matrix Z10, grain core (G1)	48	98.5 ± 0.4	1.0		260 ± 60	2670 ± 650
		SE13-123L2	НТЬ	Sumeini	in matrix Z10, grain rim (G2)	6	97.5 ± 1.5	0.1		15000 ± 7340	41 ± 18