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Subduction zones are unique to Earth and fundamental in its evolution, yet we still know little on the causes and mechanisms of their initiation. Numerical models show that far-field forcing may cause subduction initiation at weak pre-existing structures, while inferences from modern subduction zones suggest initiation through spontaneous lithospheric gravitational collapse. For both endmembers, the timing of subduction inception corresponds to initial lower plate burial, whereas coeval or delayed extension in the upper plate are diagnostic of spontaneous or forced subduction initiation, respectively. In modern systems, the earliest extension-related upper plate rocks are found in forearcs, but lower plate rocks that recorded initial burial have been subducted and are inaccessible. Here we investigate a fossil system, the archetypal Semail ophiolite of Oman, which exposes both lower and upper plate relics of incipient subduction stages. We show with Lu-Hf and U-Pb geochronology of lower and upper plate material that initial burial of the lower plate occurred before 104 Ma, predating upper plate extension and formation of Semail oceanic crust by at least 8 Myr. Such a time lag reveals far-field forced subduction initiation and

provides for the first time unequivocal, direct evidence for a subduction initiation mechanism in the geological record.

The sinking of cold lithosphere in the Earth's mantle along subduction zones is widely recognised as the main driving force for global plate tectonics[1]. Despite decades of research, the processes and mechanisms of subduction initiation remain controversial[2]. Two main conceptual end-member mechanisms considered are 'induced' and 'spontaneous' subduction initiation[2,3] (Fig. 1). Induced subduction initiation (ISI) requires a period of forced convergence, presumably accommodated at a pre-existing favourably oriented weak structure, until subduction eventually becomes self-sustained[4-6]. Alternatively, gravitational instability across oceanic transform faults or passive continental margins has been proposed to trigger lithospheric collapse and spontaneous subduction initiation (SSI) without net plate convergence[3,7]. Whether only one of ISI or SSI is the active subduction initiation mode on Earth, or both modes can be activated depending on the tectonic setting, is a matter of debate[2,3]. A fundamental criterion that would discern between ISI and SSI is the time lag between initial lower plate burial and ensuing upper plate extension (Fig. 1). During SSI, area consumed by subduction must simultaneously be balanced by area gained through upper plate extension[7,8]. In contrast, upper plate extension following ISI must be generated by the growing slab after a period of forced underthrusting[4,5,9], resulting in a time lag of several millions of years. Constraining the magnitude of this time lag requires specific geochronological methods applied to a rock record of both the formation of the incipient subduction thrust and the onset of upper plate extension.

Models for subduction initiation are based on studies of earliest extension and magmatism in the forearc of modern subductions like the Izu-Bonin-Marianna system[7,10], where rocks that directly recorded formation of the subduction interface are not exposed. Subduction initiation is widely assumed to have been spontaneous in this system[7,8,11]. Accordingly, the causes and consequences of subduction initiation are sought for in the tectonic setting at the time of forearc extension[12]. To date, the early subduction initiation history remains elusive[2], and an induced subduction initiation may have begun millions of years before forearc extension.

Research on supra-subduction zone (SSZ) ophiolites and associated metamorphic soles may yield more comprehensive insights into subduction initiation, because these provide a rock record of both upper plate extension and lower plate burial[2], respectively. SSZ ophiolites[13,14] are interpreted as relic forearc oceanic lithosphere—similar in composition to the Izu-Bonin-Marianna forearc—that formed during subduction initiation[7-9] and was subsequently uplifted above sea level[15]. Many SSZ ophiolites rest on thin (< 500m) sheets of metamorphosed oceanic crust termed metamorphic soles. These metamorphic soles derive from the uppermost crust of the subducting lower plate[16-19] that was preserved from further subduction by "welding" to the mantle section of the upper plate during subduction zone infancy[20,21]. Garnet-clinopyroxene amphibolites found at the top of many metamorphic soles indicate high-pressure granulite facies[22,23] peak metamorphic conditions (11-13 kbar and 850 °C). Metamorphism of oceanic crust to such conditions requires subduction along an anomalously hot geothermal gradient that is restricted to the initiation stage of a subduction zone[20,21,24]. These amphibolites likely represent the leading edge of the nascent slab and may therefore have directly recorded the initial burial of the lower plate during nucleation of a subduction interface[18,20,21].

The age of extension and crustal accretion in ophiolites is commonly estimated using U-Pb dating of zircon from gabbros and plagiogranites, interpreted to have formed in magma chambers below a spreading ridge[25,26]. Dating of the earliest history of the subduction interface requires estimating the age of prograde metamorphism in the garnet-clinopyroxene amphibolites of metamorphic soles. Previous chronological studies of soles used $^{40}Ar/^{39}Ar$ hornblende or mica dating[9,18,20,27,28] and more recently U-Pb dating of zircons from melt segregations[29,30]. These ages typically coincide or slightly postdate the ages of the magmatic crust of the overlying ophiolite[9,18,28]. Coinciding $^{40}Ar/^{39}Ar$ and U-Pb dates from sole rocks, and age data for ophiolitic crustal spreading have been taken to provide evidence for synchronous sole formation and upper plate spreading[28,29]. The meaning of this coincidence in terms of sole formation is nevertheless debated[9]. Both methods date post-peak conditions rather than burial[31], and thus underestimate the age of sole formation by a yet unknown amount of time.

It is clear that rigorously constraining the chronology of subduction initiation requires new approaches in dating the earliest metamorphic minerals in soles. A promising technique is Lu-Hf dating of garnet, a petrological indicator of burial and heating in metamorphosed oceanic rocks. Owing to the robustness of the chronometer at high temperatures (900-950 °C)[32-34], prograde age records of garnet growth are typically well-preserved even in cases of long-lived supra-solidus conditions[35]. Here, we apply this approach to garnet from the metamorphic sole of the archetypal Semail ophiolite of Oman to date the early stages of sole development. The results, supported by textural observations and trace element mineral chemistry, are then combined with new U-Pb zircon and titanite data, and existing dates for the sole and overlying ophiolitic crust, to investigate the complete history of the sole, from burial and heating to exhumation and cooling. The comparison of garnet growth ages in the metamorphic sole and published extension ages in the overlying ophiolite constrains a minimum time lag between initial subducting plate burial and incipient upper plate extension.

The Semail Ophiolite

The Semail ophiolite (Fig. 2a) exposes over 20,000 km^2 of oceanic crust and upper mantle rocks underlain by a discontinuous thin sheet of metamorphic sole. The ophiolite-sole couple is thought to have been emplaced in Late Cretaceous time as a giant thrust sheet[36] over the Hawasina complex comprising distal oceanic rocks, and carbonates of the Arabian passive margin[37]. The Semail ophiolite exposes a section of oceanic lithosphere including residual upper mantle rocks made of harzburgite and dunite, plutonic lower and middle crust comprising cumulates and gabbros, and an upper crustal sheeted dike complex underlying pillowed to massive submarine basalts and abyssal sediments[38] (Fig. 2). High-precision U-Pb dating of the plutonic section across the ophiolite showed that the oceanic crust of the ophiolite was generated during rapid spreading between 96.1-95.5 Ma[25,26].

The ophiolite has been classically interpreted as a relic fast spreading mid-ocean ridge[38,39]. Recent evidence, however, clearly shows that the ophiolite formed above an active subduction zone[29,40,41]. Similarities in the chemostratigraphy of the Semail ophiolite and the Philippine Sea Plate forearc strengthened the inference that the ophiolite formed during subduction initiation[2,13], either spontaneous[3,8,42] or induced[43,44].

The sole of the Semail ophiolite comprises amphibolites that are notably garnet- and clinopyroxene-bearing near the contact with mantle rocks. Garnet-clinopyroxene sole amphibolites represent oceanic upper crustal MORB-like basaltic sequences[40] of unknown age, which were metamorphosed to peak conditions of 11-13 kbar and 850 °C[21,45,46]. U-Pb dating of zircon from melt segregations suggests solidification of the melt fraction by 96.16-94.5 Ma[26,29,30], followed by rapid cooling below the closure temperature of the $^{40}Ar/^{39}Ar$ system in hornblende (500-550°C) between 95.7 and 92.6 Ma[28]. The available data support the hypothesis that the upper crustal protolith of the sole was subducted to mantle depths in excess of 35 km along an incipient hot subduction plane before being transferred to the upper plate[20,21].

This study focuses on two main Omani metamorphic sole localities: Wadi Tayin and Wadi Sumeini (Fig. 2). The Wadi Tayin locality[21,45] (Fig. 2b) exposes amphibolites interlayered with thin quartz- and calc-silicate-rich layers, overlain by a middle quartzite dominated interval, and again an amphibolite layer with garnet amphibolites present in the top 5 m (Fig. 2e). The Sumeini sole locality[45,46] (Fig. 2c) also exposes amphibolites, which become garnet- and clinopyroxene-bearing in the upper 10 m (Fig. 2d), whereas the lower section of the sole consists of epidote amphibolite with more abundant quartzite and marble. We collected samples WT-150 and WT-151 from the Wadi Tayin and SU-03A from the Sumeini sole localities (Figs. 2b, 2c, 2d, 2e) from garnet- and clinopyroxene-bearing amphibolites that occur as meter-scale coherent levels (Fig. 2d) or as boudins embedded in garnet-free amphibolite (Fig. 2e) immediately below the contact with the overlying mantle section.

Occurrence, composition and age of garnet, zircon and titanite

The samples show a hornblende-dominated nematoblastic fabric that wraps around boudinaged bands of garnet-clinopyroxene-rich granulite (Figs. 3a, 3b, 3c). Garnet occurs as subhedral cm-scale porphyroblasts with abundant inclusions (Fig. 3). Mineral compositions are consistent with those of similar samples used in previous petrological studies[21,45,46] and with high-pressure granulite facies metamorphic conditions[22,23]. The strongly foliated matrix is defined by subhedral hornblende and subordinate anhedral diopside with abundant pseudomorphed anhedral plagioclase and fine-grained ilmenite-titanite symplectites. The granulite assemblage is variably overprinted by dynamic amphibolite-facies metamorphism. Sample SU-03A best preserves the granulite assemblage, whereas sample WT-150 shows the strongest amphibolite overprint. A lower-grade assemblage with epidote, prehnite and albite is found in fractures and veins, and as pseudomorphic replacements of plagioclase; the mafic minerals of the granulite assemblages are not substantially affected by such replacement.

Garnet shows complex zoning and inclusion patterns that differ between samples. Two garnet zones (grt-1 and grt-2) are nevertheless consistently observed. Grt-1 is defined by anhedral cores (grt-1) that are generally rich in Ca, Mn, and HREE (Figs. 3, 4). These cores are typically poikiloblastic with inclusions of titanite and apatite in innermost domains (grt-1a) and mono- and polymineralic inclusions of diopside, hornblende, plagioclase, ilmenite, titanite, and apatite in outer domains (grt-1b). Polymineralic inclusions locally show negative shapes and very low dihedral angles (Figs. 3f, 3g), suggesting they represent the solidification product of trapped melt. Grt-1a shows distinctly lower chondrite-normalised Gd/Yb than grt-1b. Grt-2 is defined by a textural and compositional mantle (grt-2) that encloses anhedral grt-1 cores. Grt-2 has fewer inclusions, is Mg-rich and Ca-poor, has high Gd_N/Yb_N, and shows strong and locally very well-defined oscillatory zoning for HREE (Fig. 4).

All three samples yielded garnet-whole rock Lu-Hf isochrons (Figs. 5a, 5b, 5c) with MSWD between 0.32 and 0.79, and uncertainties of 0.8 %RSD or better. The samples from Wadi Tayin yielded 104.1 ± 1.1 Ma (MSWD = 0.79; 150A) and 103.2 ± 1.2 Ma (MSWD = 0.32; 151A), and sample SU-03A from Sumeini provided 103.5 ± 1.6 Ma (MSWD = 0.62). All Lu-Hf age data are identical within uncertainty. Taking a weighted mean of these ages yields 103.7 ± 0.7 Ma (MSWD = 0.63), indicating no resolvable age scatter among the samples.

Zircon and titanite grains were recovered from sample WT-151. The zircon population consists largely of colourless, subequant and anhedral grains (Fig. 5d). Five analyses of such grains are clustered to the right of the Concordia curve. The slight discordance and spread in $^{207}Pb/^{235}U$ is a common feature of young zircon populations reflecting in part the likely bias in the decay constants used and initial ^{231}Pa excess[47]. All five $^{206}Pb/^{238}U$ ages overlap within error yielding a robust average age of 96.19 ± 0.14 Ma for crystallization of zircon. A fraction of four titanite grains provides a $^{206}Pb/^{238}U$ age of 95.60 ± 0.27 Ma.

Garnet growth in the Omani ophiolite soles

The microtextures and major and trace element distributions described above indicate the following growth history (Fig. 6). Grt-1 nucleated and initially grew at subsolidus conditions in the titanite stability field. The transition from grt-1a to grt-1b marks the prograde stabilization of ilmenite and the formation of inclusions indicative of the first occurrence of melt. Peritectic garnet growth at those conditions is further supported by evidence of melt segregations at the outcrop scale[29,30] and by phase equilibria modelling[21,23]. The distinct increase in Gd_N/Yb_N could relate to the dehydration melting of hornblende or to titanite break-down. Grt-2 represents peritectic garnet as indicated by its oscillatory zoning. This zoning is interpreted to reflect the competition between the rates of HREE uptake by growing garnet and diffusive HREE supply within the melt. Such a garnet growth sequence is consistent with phase equilibria modelling and experiments for MORB-like protoliths that predict suprasolidus grt-2 growth[21,23] from 9 kbar-650°C to 11 kbar-850°C across the titanite-ilmenite transition[48].

The robustness of the Lu-Hf geochronometer[33,34] is largely governed by the low diffusivity of Hf[49]. Closure temperatures of diffusive Hf loss for the grains analyzed are at least 900 °C[33] and hence exceed peak temperatures that the Omani sole samples were subjected to. Lu is more mobile and modelled mechanisms of age skewing by diffusive Lu redistribution[49] must be considered. These, however, clearly are not applicable here. The dated samples show exceptional preservation of the fine growth zoning in the distributions of Lu, which precludes any significant diffusive homogenization of Lu after garnet growth. The dates, which were determined for bulk-grain garnet populations, therefore represent an estimate of the average age of garnet growth weighted according to Lu distribution. The Lu-Hf dates for all three samples are identical, yet show different Lu distributions. This shows that weighing of ages was insignificant. Our data are thus best explained by a single, fast garnet growth event at sub- to suprasolidus conditions from roughly 550°C and 8 kbar to peak conditions of 850°C and 11-13 kbar (Fig. 6). Zircon (96.19 ±

0.14 Ma) may have crystallized from late highly fractionated solidifying trapped melt[29,31] when the rocks cooled from peak conditions to subsolidus conditions of 700°C. The U-Pb dates of titanite (95.60 ± 0.27 Ma) represent cooling below 650 - 600°C[31].

Implications for subduction initiation

Rheological studies indicate that the upper part of a subducting oceanic plate will be transferred to the upper plate when conditions of 850°C and 11-13 kbar are reached at the interface, forming a metamorphic sole[20,21]. We now show that garnet growth in the sole under the Semail ophiolite occurred at 104 Ma. This age provides a timing for burial, decoupling from the lower plate, and transfer to the upper plate. Between peritectic garnet growth at 104 Ma and zircon crystallization at 96 Ma, the welded sole did not record any major thermal or dynamic perturbation, likely staying at supra-solidus peak conditions while underthrusting progressed. Around 96 Ma, extension in the upper plate leads to oceanic lithosphere accretion along a spreading centre. From 96.2 to 94.5 Ma, zircon crystallized from the segregated melt fractions[29,30] in the underlying metamorphic sole, marking cooling to subsolidus conditions[29,31] from >850°C to ~700°C. Cooling to 600-650°C[31] occurred ~0.5 Myr later as shown by our titanite U-Pb age, whereas cooling to ~550°C and below is constrained by hornblende $^{40}Ar/^{39}Ar$ dating between ~95.5 Ma to ~92 Ma[28]. The onset of cooling in the sole thus coincides with the formation of SSZ oceanic crust (96.12-95.50 Ma[25,26]). In the Semail ophiolite, sole formation, or lower plate burial, started >8 Myr before upper plate extension occurred. The inference that underthrusting below the mantle section predated formation of the ophiolitic crust by at least 8 Myr confirms a SSZ origin for the Semail ophiolitic crust, settling the long discussion regarding its origin [29,38,39,41,50].

The Semail ophiolite, preserving a ~50 km wide forearc lithosphere[8] measured perpendicular to its spreading direction[38], is 8 Myr younger than the sole age, and thus does not preserve the crust of the pre-subduction initiation lithosphere. Therefore, we cannot conclude with certainty that this is the oldest SSZ crust that formed after subduction initiation. However, during SSZ spreading, the ridge must have moved away from the trench at half-spreading rate[9], which was >10cm/yr[25]. At these rates, if upper plate spreading had started even a million years earlier, the ophiolite should have been ~50-100 km wider than today to preserve the 96-95.5 Ma old crust. We therefore conclude that the oldest crust of the Semail ophiolite formed at the onset of upper plate extension.

Garnet growth in the metamorphic sole at 104 Ma and onset of SSZ crustal accretion by 96 Ma constrains a >8 Myr time lag between initial lower plate burial and the onset of upper plate extension, implying a >8 Myr period of forced convergence prior to upper plate extension. This time lag constitutes the first direct evidence from the geological record for ISI. The far-field causes driving the forced convergence should be sought considering the pre-104 Ma plate configuration and evolution, not the 96 Ma syn-upper plate extension configuration.

Our new results imply that SSZ ophiolite formation is not unequivocal evidence for SSI, as often assumed for Semail and other large and well-preserved ophiolites[3,8,42]. In fact, SSZ ophiolite formation rather indicates the onset of upper plate extension, which does not date subduction initiation in ISI.

Both the magnitude of the time lag between initial lower plate burial and incipient upper plate extension, and the age of the onset of convergence convey critical, previously unavailable information on subduction initiation. The magnitude of the time lag should reflect the balance between forces driving and resisting upper plate extension, depending on the nature, geometry and kinematics of the intervening plates, as indicated by numerical models[2,4-6]. Longer time lags could indicate a strong upper plate, a long subduction interface or a slow convergence rate. Models of ISI at transform faults involving a very young upper plate[4-6] predict time lags of the order of 5-10 Myr, corresponding very well to our results. Nonetheless, the tectonic setting that led to initiation of subduction and formation of the Semail ophiolite must be validated from the rock record, in the pre-104 Ma configuration. Accordingly, the absolute timing of initial lower plate burial is also of utmost importance. The plate configuration and kinematics in which new subduction zones were initiated in the geological past might have significantly predated the earliest expression of upper plate spreading represented by ophiolitic or modern forearc crust. Such new insights into subduction initiation processes open new avenues for reconceptualization of the initiation and processes of global plate tectonics.

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Author contributions

C.G. generated the project, led field work, completed the petrological study and wrote the manuscript.

M.S. conducted the Lu-Hf analyses and contributed to writing the manuscript

D.V.H. participated in field work, contributed to the rationale and to writing the manuscript

D.G. and F.C. completed and treated the U-Pb geochronological analyses

B.C. planned and participated in field work, prepared and analyzed samples

M.M. organised and participated in field work

O.R. participated in defining the rationale and writing the manuscript

D.S. conducted and treated the LA-ICP analyses

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We have no competing financial interests.

Figure Captions

Figure 1 : Conceptual lithospheric sections representing spontaneous vs. induced subduction initiation. The time lag between initial lower plate burial and incipient upper plate extension is diagnostic of subduction initiation mode.

Figure 2: Geological maps, sample locations and field relationships. a) Geological map of Oman, modified from Nicolas et al. 2000 and Rioux et al. 2016. b) and c) Geological maps and sample locations at Wadi Tayin and Sumeini, maps after Cowan et al. 2014. d) and e) field relationships for samples SU-03 and WT-151. The scale bar on the peridotite outcrop of e) is 10 cm.

Figure 3: Petrography of the investigated samples. a), b) and c) are micro-XRF chemical maps of thin sections for samples SU-03A, WT-150 and WT-151. d) and e) are close ups. f) and g) are BSE images of a melt pseudomorph inclusion in garnet from SU-03A. h), i) and j) are EPMA chemical profiles (located of 3a, 3b and 3c) showing almandine (Alm), grossular (Gr), pyrope (Py) and spessartine (Sp) mole fractions and Mg# = $Mg^{2+}/(Mg^{2+} + Fe^{2+})$.

Figure 4: Trace element content of representative garnet from samples SU-03A, WT-150 and WT-151. a), c) and e) Lu maps, location of maps is shown in figure 3a, 3b and 3c. b), d) and f) REE profiles normalised to chondrites (Sun and McDonough, 1989). Location of spot analyses is shown in 3a, 3c and 3e.

Figure 5: Geochronological results. a), c) and c) isochrons for garnet and whole rock fractions of samples SU-03A, WT-150 and WT-151. Data is available in table 3. d) Concordia diagram for

zircon and titanite of sample WT-151. Zircon grains under binocular in 1 x 1 mm inset. Data is available in table 4.

Figure 6: P-T-t evolution of the Semail metamorphic sole. P-T trajectory is from Soret et al. 2017. Supra-solidus garnet growth ages are from this study. Zircon ages are from this study and from Rioux et al. 2016. Titanite ages are from this study. Hornblende ages are compiled in Soret et al. 2017. 10% partial melting isomodes and solidus for MORB-like protoliths are from Palin et al. 2016. Garnet-in boundary and titanite-ilmenite transition are from Liu et al. 1996. I, II and III are lithospheric section diagrams synthesizing our results. In III, spreading ages are from Rioux et al. 2012; 2013. See text for explanations.

Methods

Representative thin sections for each sample were mapped using a Bruker M4 Tornado μ-XRF instrument at Université Laval (Figure 3a, 3b, 3c) equipped with two 60 mm^2 Silicon Drift Detectors, operating at 50 kv and 300 nA with a step size of 20 μm and a dwell time of 3 ms per pixel, to find garnet grain sections that intersected the core. These garnet grains were subjected to major-element quantitative point analysis along radial profiles using a Cameca SX-100 five sprectrometer electron probe microanalyzer at Université Laval. Analytical conditions were 15 kV, 20 nA with a counting time of 20 s on peaks and 10 s on background. Calibration standards used were generally simple oxides (GEO Standard Block of P&H Developments), or minerals where needed (Mineral Standard Mount MINM25-53 of Astimex Scientific Limited; reference samples from[51]). Data were reduced using the PAP model. The data are available in Table 1 and figures 3h, 3i and 3j.

Trace-element analysis of the garnet sections was done by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at LabMaTer (Université du Québec à Chicoutimi), using a RESOlution 193 nm excimer laser (Australian Scientific Instrument) and a S155 Laurin Technic ablation cell system coupled to an Agilent 7900 quadrupole ICP-MS. Spot analyses were conducted with a 33 µm beam operating at 15 Hz, 5 J/cm^2 in a 4ms/isotope cycle. High resolution mapping was done with a 20 µm beam at a speed of 80 µm/s (figures 4a, 4c and 4e) and pulsing of 30 Hz at $5J/cm^2$ in a 4 ms/isotope cycle. Calibrant used was the synthetic basalt glass GSE-1G (USGS), using preferred values from the GEOREM database[52]. The data have been processed with IOLITE freeware[53] to generate maps and achieve fully-quantitative results on spots analysis using ^{29}Si as internal standard. The data are available in Table 2 and figure 4.

For garnet Lu-Hf and zircon U-Pb geochronology, samples were disaggregated using an Electric Pulse Disaggregation instrument at Overburden Drilling Management Ltd to 90% <1mm. Bulk-rock powders were created from this fraction. Large garnet concentrates of 800 mg or more were extracted from the samples through standard concentration methods: sieving, magnetic separation using a Frantz magnetic barier separator, heavy liquor density separation and hand-picking on a binocular microscope. Zircon and titanite grains were handpicked from the heavy mineral fraction.
Garnet Lu-Hf chronology was done at the Pacific Centre for Isotopic and Geochemical Research, University of British Columbia. There, garnet crystals and bulk-rock powder were transferred to

screw-top PFA vials and weighed. Garnet grains were then washed using de-ionized water and bathed in 1 N HCl at room temperature for 1 h. After removing the HCl, garnet samples were dried, mixed with a ^{176}Lu- ^{180}Hf isotope tracer that has a Lu/Hf similar to that of generic garnet, and digested through repeated addition of HF:HNO_3:$HClO_4$ and 6 N HCl, each step followed by evaporation to dryness. After admixing of a mixed ^{176}Lu- ^{180}Hf isotope tracer with low Lu/Hf, the bulk-rock powders were digested in a stainless-steel digestion vessel at 180 °C for 7 days using HF:HNO_3.

After digestion, all samples were dried down, re-dissolved in 6 N HCl, diluted to 3 N HCl using de-ionized H_2O, and centrifuged. The solution containing the garnet elemental solute was then loaded onto polypropylene columns containing a 1-ml Ln-Spec® resin bed and subjected to REE-HFSE chromatography modified from the method of Münker et al.[54]. Isotope analyses for Hf and Lu were done using the Nu Instruments *Plasma HR* multi-collector (MC) ICPMS at PCIGR. For Lu analyses, isobaric interference of ^{176}Yb on m/z corresponding to mass 176 was corrected using an exponential correlation between ^{176}Yb/^{171}Yb and ^{174}Yb/^{171}Yb. This correlation was calibrated through replicate analyses of Yb solution standards from the National Institute of Standards and Technology performed at different concentrations (10-100 ppb[55]). For Hf isotope analyses, ^{180}Ta and ^{180}W interferences were estimated on the basis of ^{181}Ta/^{177}Hf and ^{183}W/^{177}Hf, assuming natural abundance and a Hf-based mass bias. Mass bias was assumed to follow an exponential law and was corrected for applying ^{179}Hf/^{177}Hf = 0.7325 (Hf, Ta, W) and ^{173}Yb/^{171}Yb = 1.1296 (Lu, Yb). Any resolvable drift was corrected for assuming linear time dependence. Hafnium isotope ratios are reported relative to the JMC-475 Hf standard (^{176}Hf/^{177}Hf = 0.28216[56]). The external ^{176}Hf/^{177}Hf reproducibility (2 s.d.) of replicate JMC-475 analyses done at concentrations similar to those of sample solutions was 0.4 ε_{Hf} during the course of our analytical sessions. The external reproducibility of $^{176}Hf/^{177}Hf$ was estimated from the standard scatter at the given sample concentration and internal error. This estimate was made by comparing internal and external uncertainty for replicate analyses of JMC-475 done at concentrations that bracketed those of samples (10-50 ppb[57]). The Lu-Hf isochrons were established using *Isoplot* v. 3.27[58] applying 1.876×10^{-11} yr^{-1} for $\lambda^{176}Lu$[59,60]. All uncertainties are cited at the 2-s.d. level. The results are provided in Table 3 and Figure 5.

The samples were screened for zircon and titanite; both minerals were found only in sample 151A. After selection under an optical microscope zircon was subjected to chemical abrasion[61,62] whereas titanite was not abraded. The selected grains were then spiked with a ^{202}Pb-^{205}Pb-^{235}U tracer, followed by dissolution, chemical separation of Pb and U, and mass spectrometry, after the procedure detailed in Krogh[63] with modifications described in Corfu[64]. The Pb measurements were done mostly with an ion counting secondary electron multiplier. The obtained data were corrected with fractionation factors determined from the 205Pb/202Pb ratio of the tracer (around 0.1%/amu for Pb) and 0.12%/amu for U, subtracting blanks of 0.1 pg U and 2 pg Pb, or less when the total common Pb was below that level. The remaining initial Pb was corrected using compositions calculated with the model of Stacey and Kramers[65]. The data were also adjusted for a deficit of ^{206}Pb due to initial deficiency of ^{230}Th[66] and the tracer was calibrated with reference to the ET100 solution (Condon, personal communication, 2014). Plotting and regressions were done with the Isoplot software package[58]. The decay constants are those of Jaffey et al.[67]. The results are provided in Table 4 and figure 5.

The authors declare that all the data supporting the findings of this study are available within the paper and its supplementary information files.

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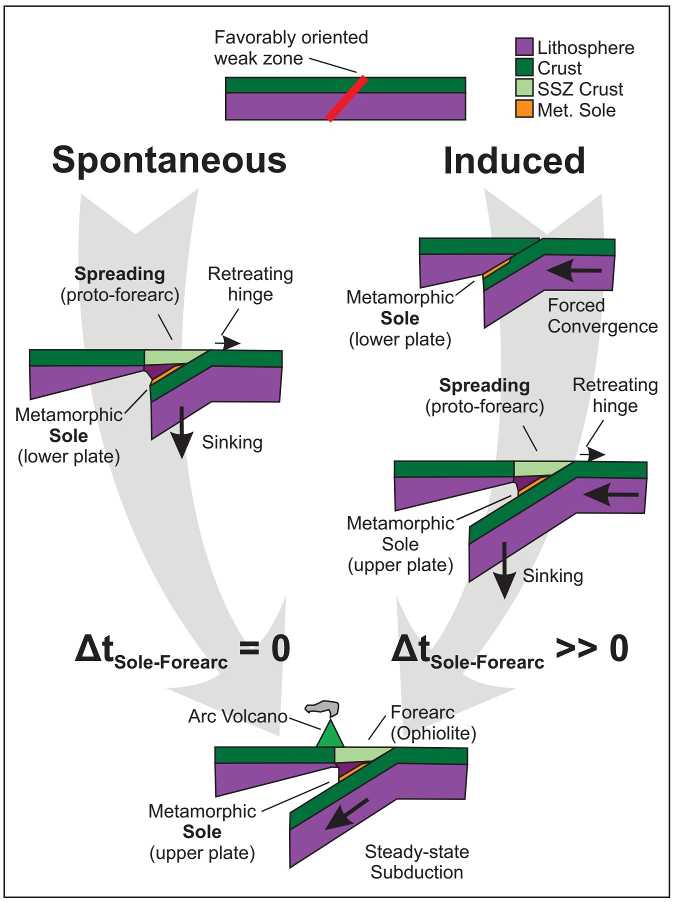
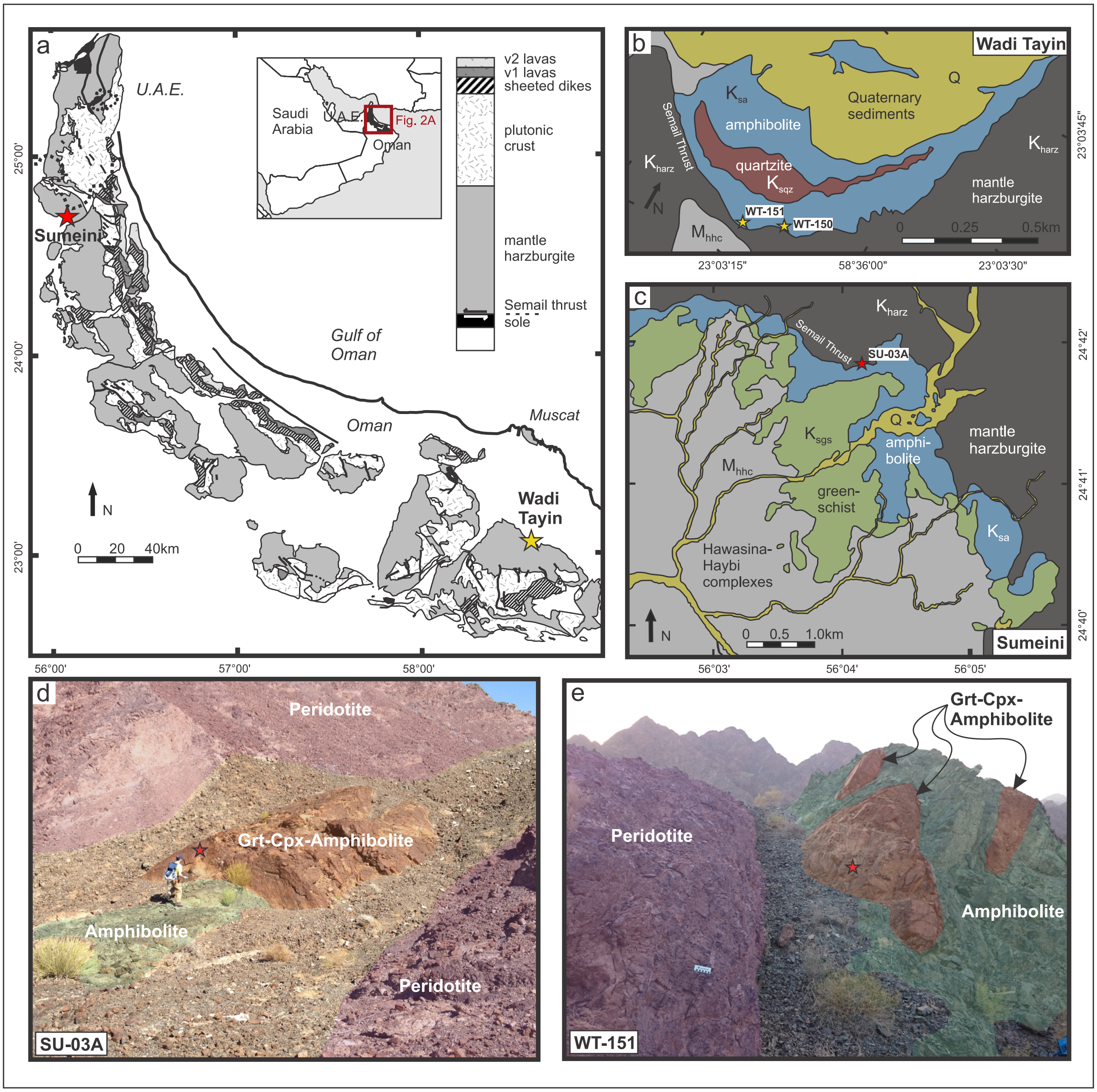
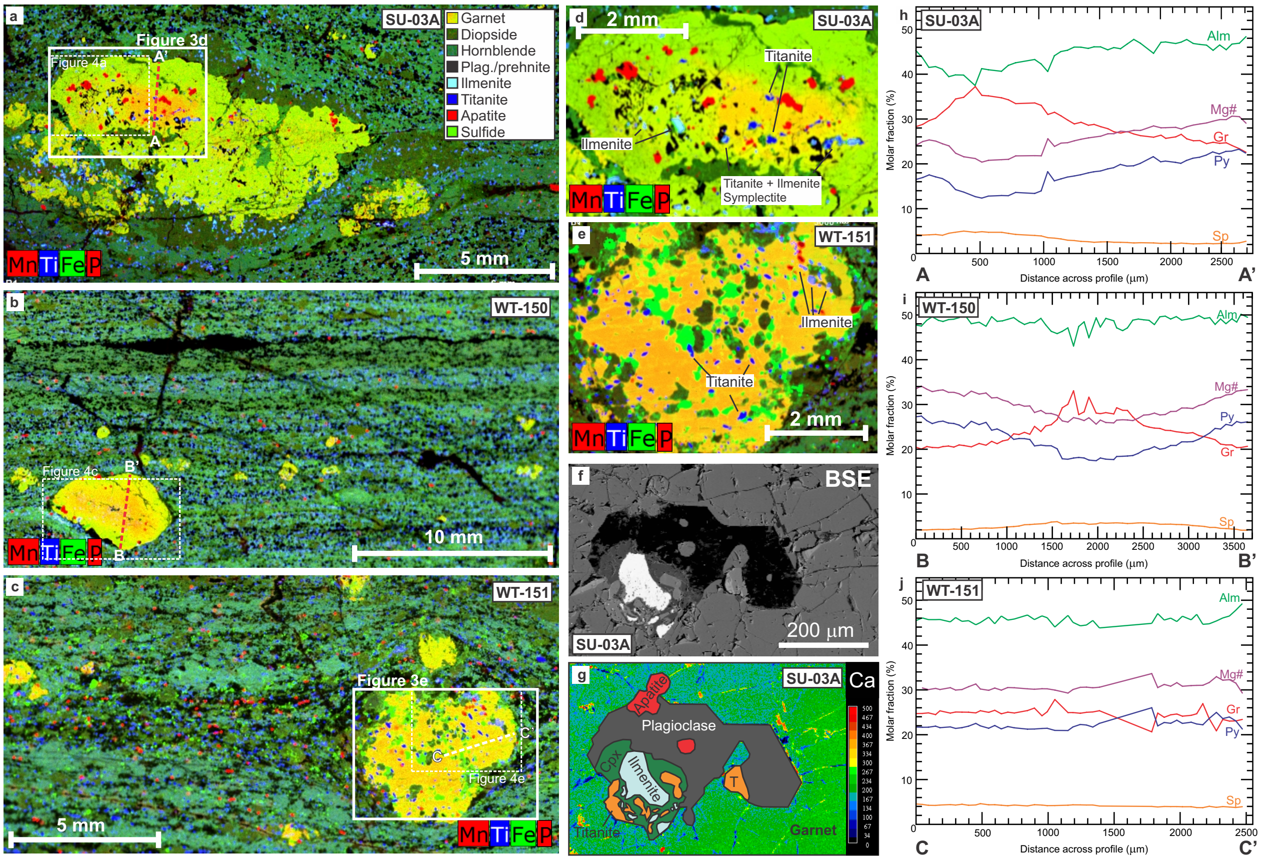
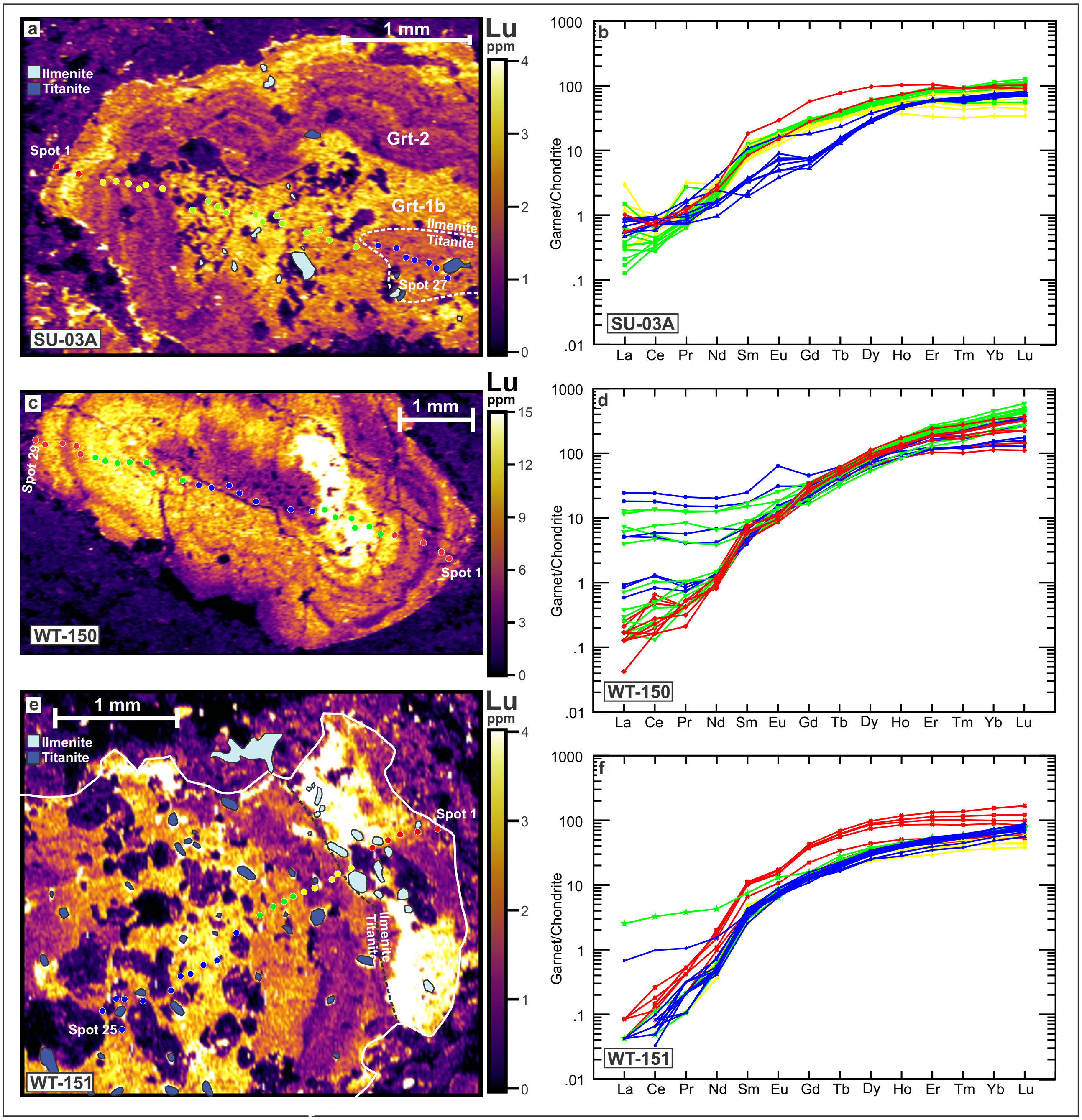


Favorably oriented weak zone
Lithosphere
Crust
SSZ Crust
Met. Sole
Spontaneous
Induced
Spreading (proto-forearc)
Retreating hinge
Metamorphic Sole (lower plate)
Sinking
Metamorphic Sole (lower plate)
Forced Convergence
Spreading (proto-forearc)
Retreating hinge
Metamorphic Sole (upper plate)
Sinking
$\Delta t_{Sole\text{-}Forearc} = 0$
$\Delta t_{Sole\text{-}Forearc} >> 0$
Arc Volcano
Forearc (Ophiolite)
Metamorphic Sole (upper plate)
Steady-state Subduction

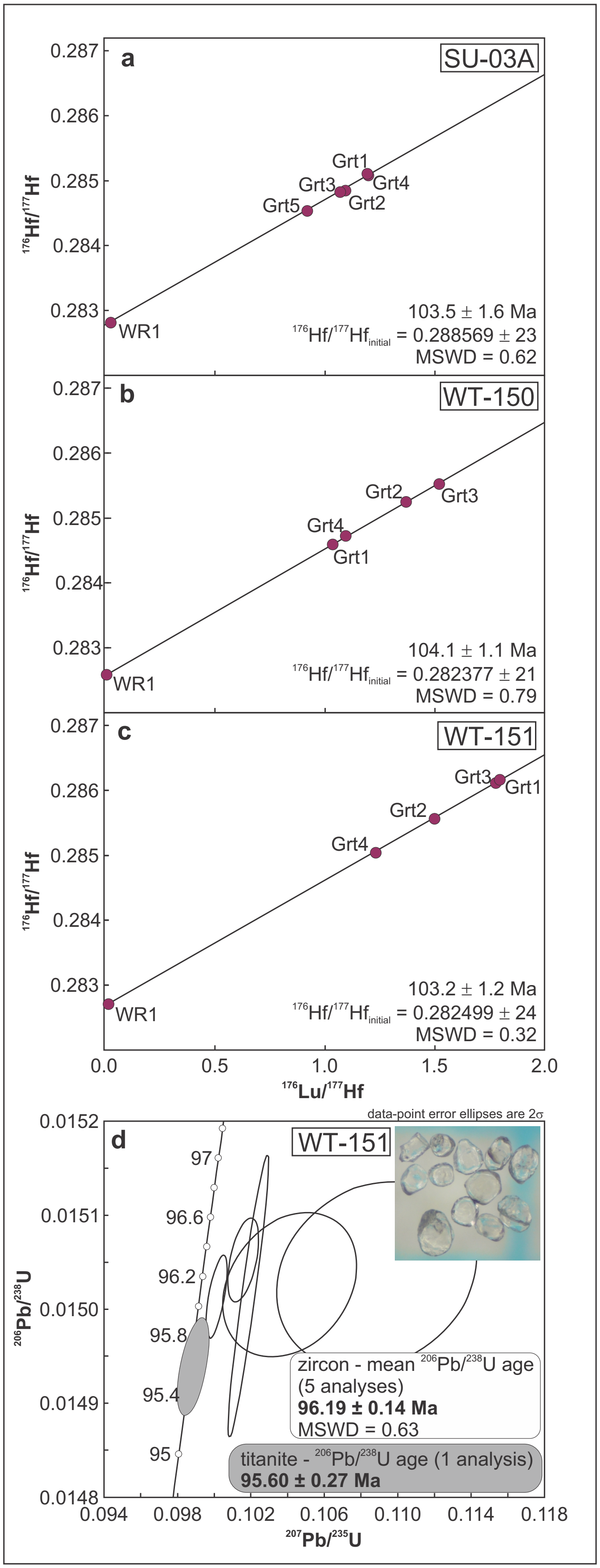

a
U.A.E.
Saudi Arabia
U.A.E.
Oman
Fig. 2A
v2 lavas
v1 lavas
sheeted dikes
plutonic crust
mantle harzburgite
Semail thrust sole
Sumeini
Gulf of Oman
Oman
Muscat
Wadi Tayin
N
0 20 40km
25°00'
24°00'
23°00'
56°00'
57°00'
58°00'
b
Wadi Tayin
Semail Thrust
K_harz
K_sa
amphibolite
Quaternary sediments
Q
quartzite
K_sqz
mantle harzburgite
K_harz
N
M_hhc
WT-151
WT-150
0 0.25 0.5km
23°03'15"
58°36'00"
23°03'30"
23°03'45"
c
K_harz
Semail Thrust
SU-03A
K_sgs
Q
amphi-bolite
mantle harzburgite
M_hhc
green-schist
K_sa
Hawasina-Haybi complexes
N
0 0.5 1.0km
Sumeini
56°03'
56°04'
56°05'
24°42'
24°41'
24°40'
d
Peridotite
Grt-Cpx-Amphibolite
Amphibolite
Peridotite
SU-03A
e
Grt-Cpx-Amphibolite
Peridotite
Amphibolite
WT-151

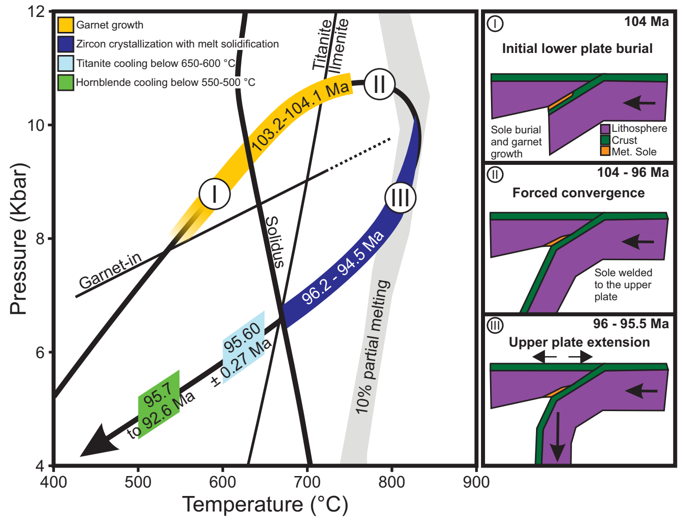


a
SU-03A
Figure 3d
Figure 4a
A'
A
Garnet
Diopside
Hornblende
Plag./prehnite
Ilmenite
Titanite
Apatite
Sulfide
Mn Ti Fe P
5 mm
b
WT-150
Figure 4c
B'
B
10 mm
c
WT-151
Figure 3e
C'
C
Figure 4e
5 mm
d
2 mm
SU-03A
Titanite
Ilmenite
Titanite + Ilmenite Symplectite
e
WT-151
Ilmenite
Titanite
2 mm
f
BSE
200 μm
SU-03A
g
SU-03A
Ca
Apatite
Plagioclase
Cpx
Ilmenite
T
Titanite
Garnet
h
SU-03A
Alm
Mg#
Gr
Py
Sp
Molar fraction (%)
Distance across profile (μm)
i
WT-150
j
WT-151



a
1 mm
Lu ppm
Ilmenite
Titanite
Grt-2
Spot 1
Grt-1b
Ilmenite
Titanite
Spot 27
SU-03A
b
Garnet/Chondrite
SU-03A
La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
c
1 mm
Spot 29
Spot 1
WT-150
d
WT-150
e
1 mm
Ilmenite
Titanite
Spot 1
Spot 25
Ilmenite
Titanite
WT-151
f
WT-151


a
SU-03A
Grt1
Grt3
Grt4
Grt2
Grt5
WR1
103.5 ± 1.6 Ma
$^{176}Hf/^{177}Hf_{initial}$ = 0.288569 ± 23
MSWD = 0.62
b
WT-150
Grt2
Grt3
Grt4
Grt1
WR1
104.1 ± 1.1 Ma
$^{176}Hf/^{177}Hf_{initial}$ = 0.282377 ± 21
MSWD = 0.79
c
WT-151
Grt3
Grt1
Grt2
Grt4
WR1
103.2 ± 1.2 Ma
$^{176}Hf/^{177}Hf_{initial}$ = 0.282499 ± 24
MSWD = 0.32
$^{176}Hf/^{177}Hf$
0.283
0.284
0.285
0.286
0.287
$^{176}Lu/^{177}Hf$
0.0
0.5
1.0
1.5
2.0
data-point error ellipses are 2σ
d
WT-151
97
96.6
96.2
95.8
95.4
95
zircon - mean $^{206}Pb/^{238}U$ age
(5 analyses)
96.19 ± 0.14 Ma
MSWD = 0.63
titanite - $^{206}Pb/^{238}U$ age (1 analysis)
95.60 ± 0.27 Ma
$^{206}Pb/^{238}U$
0.0148
0.0149
0.0150
0.0151
0.0152
$^{207}Pb/^{235}U$
0.094
0.098
0.102
0.106
0.110
0.114
0.118



Garnet growth
Zircon crystallization with melt solidification
Titanite cooling below 650-600 °C
Hornblende cooling below 550-500 °C
103.2-104.1 Ma
96.2 - 94.5 Ma
95.60
± 0.27 Ma
95.7
to 92.6 Ma
Garnet-in
Solidus
Titanite
Ilmenite
10% partial melting
I
II
III
Pressure (Kbar)
Temperature (°C)
12
10
8
6
4
400
500
600
700
800
900
I 104 Ma
Initial lower plate burial
Sole burial and garnet growth
Lithosphere
Crust
Met. Sole
II 104 - 96 Ma
Forced convergence
Sole welded to the upper plate
III 96 - 95.5 Ma
Upper plate extension