



Research Paper

Revisiting the Arabian-Nubian shield-Saharan meta-craton boundary: Evidence for a ca. 840–655 Ma Keraf intra-oceanic arc terrane

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ABSTRACT

The arc-continent suture zone separating the Neoproterozoic Arabian-Nubian shield from Archean-Paleoproterozoic crust of the Saharan meta-craton has long been thought to coincide with the Keraf structural corridor. This interpretation remains debated, and it is unclear whether crustal blocks on either side of the hypothesised suture have distinct geological histories. In this study, we integrate field observations, whole-rock geochemistry, zircon U-Pb-Hf and whole-rock Sr-Nd isotopic data from plutons sampled along an ~ 80 km-long transect across the central Keraf domain and propose that it represents a structurally modified, juvenile intra-oceanic arc terrane formed between ca. 840–655 Ma.

Pre- to syn-tectonic intrusions define a two-stage intra-oceanic arc history: (i) juvenile, mantle-derived arc magmatism between ca. 840–810 Ma ($\epsilon_{\text{Hf}(t)} \sim +9.1$ to $+10.7$, $\epsilon_{\text{Nd}(t)} \sim +3.8$ to $+4.7$), and (ii) protracted arc thickening by magmatic accretion between ca. 760–655 Ma ($\epsilon_{\text{Hf}(t)} \sim +7.7$ to $+9.0$, $\epsilon_{\text{Nd}(t)} \sim +3.7$ to $+6.2$), with contribution from early arc crust material. The apparent absence of zircon crystallisation ages between ca. 810–760 Ma may reflect trenchward arc migration, although alternative processes cannot yet be excluded, and the significance of this pattern remains unclear. Post-collisional granites at ca. 620 Ma ($\epsilon_{\text{Hf}(t)} \sim +8.2$ to $+8.8$; $\epsilon_{\text{Nd}(t)} \sim +3.3$ to $+4.6$) likely record reworking of the Keraf arc crust, with limited, if any, enriched-mantle input during the late to post-collisional period.

Integration of our results with regional isotopic compilations shows that the isotopically more evolved Bayudian (ca. 970–890 Ma) and syn-Keraf arc magmatic events affected the pre-Neoproterozoic basement in the easternmost Saharan meta-craton. Although this hypothesis remains to be demonstrated, the available data are consistent with a scenario involving subduction transference from the Bayuda continental arc to the Keraf intra-oceanic arc at ca. 890–860 Ma, implying that the Arabian-Nubian shield-Saharan meta-craton suture zone may lie farther west than currently accepted.

1. Introduction

Terrane analysis in ancient orogens can be challenging, as different authors have proposed alternative boundaries between basement terranes (as defined by Howell, 1985), due to limited constraints on the age and nature of stratigraphic, magmatic, metamorphic, and tectonic events in adjacent geological domains. Notably, the position of the boundary between West Gondwana to the west and the Neoproterozoic East Africa-Antarctic Orogen (EAAO) to the east remains highly debated

for its northern extent, along the Neoproterozoic Arabian-Nubian shield (ANS; Fig. 1a, b).

In the Nubian shield, a long-standing view is that the arc-continent suture zone between the juvenile Neoproterozoic ANS to the east and Archean-Paleoproterozoic crust of the Sahara meta-craton to the west lies within the Keraf domain, east of the Bayuda domain (Fig. 1b; Abdelsalam et al., 1998; Bailo et al., 2003; Fritz et al., 2013; Abdelrahman et al., 2017). Under this configuration, the Atmur-Delgo suture zone, developed north and northwest to the Keraf domain, has been

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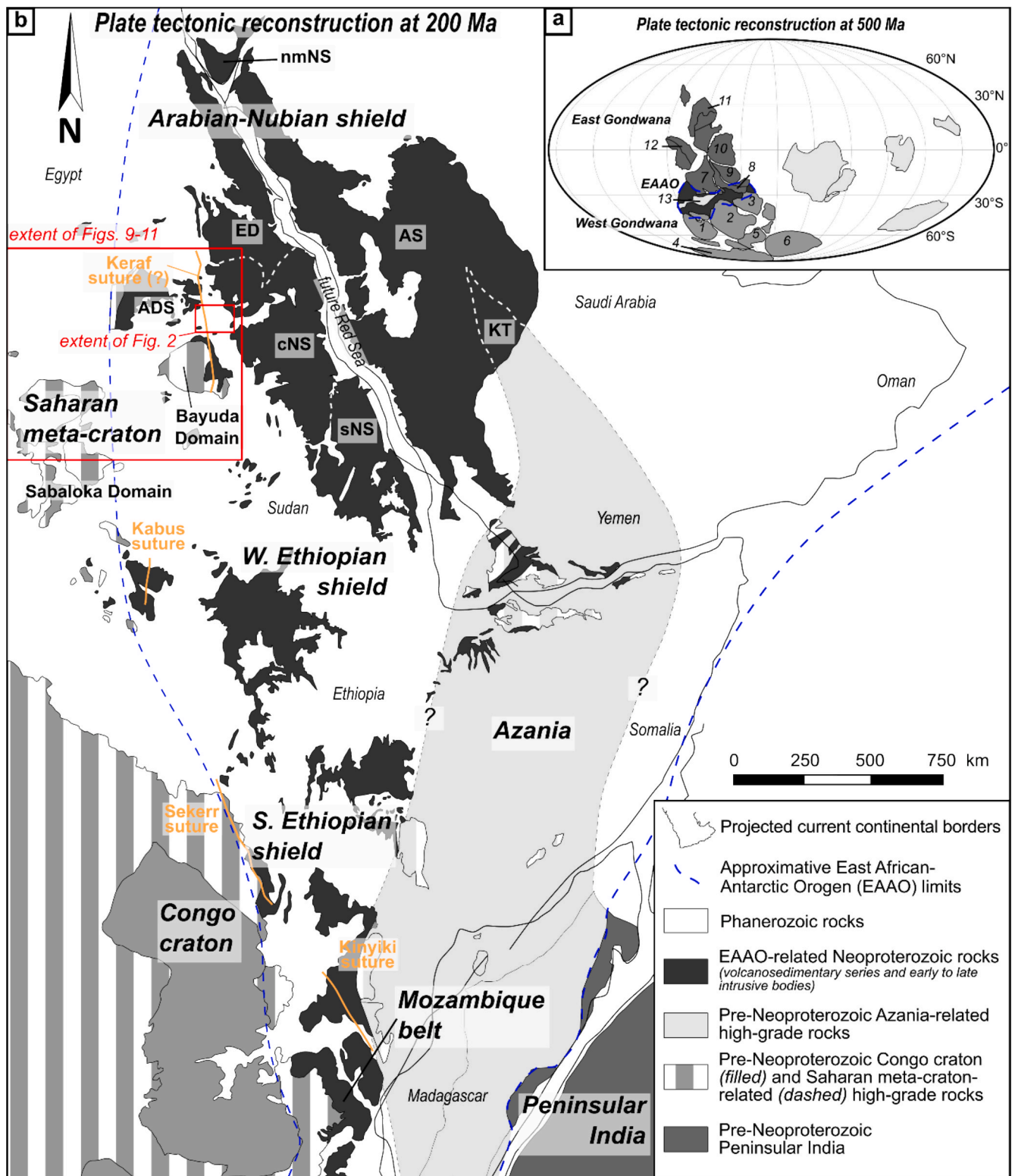


Fig. 1. a. plate tectonic reconstruction at 500 Ma showing the ca. 950–550 Ma East African-Antarctic Orogen (EAAO) between East and West Gondwana, modified after Grenholm (2019) and references therein (Johnson et al., 2011; Fritz et al., 2013; Johnson, 2014; Merdith et al., 2017). Numbered cratons and crustal blocks are 1: Saharan meta-craton; 2: Congo craton; 3: Kalahari craton; 4: West African craton; 5: Sao Francisco craton; 6: Amazon craton; 7: Peninsular India; 8: Rayner Province; 9: Chron craton; 10: Terre Adélie; 11: Australian cratons; 12: South China craton; 13: Azania. b. Regional map of the EAAO prior to Gondwana break-up at ca. 200 Ma, simplified after Grenholm (2019). The red rectangles delimit the extent of (i) Fig. 2 across the Kerat domain in the western Arabian-Nubian shield, and (ii) Figs. 9–11 across the boundary between the Nubian shield and the Saharan meta-craton. The Arabian-Nubian shield subdivisions are ADS: Atmur-Delgo suture; AS: Arabian shield minor the Khida terrane; cNS: central Nubian shield; ED: Eastern Desert; KT: Khida terrane, isotopically distinct from the rest of the Arabian shield; nmNS: northernmost Nubian shield (Sinai Peninsula); sNS: southern Nubian shield. Extents of the Kerat, Kabus, Sekerr, and Kinyiki sutures in orange are after Fritz et al. (2013). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

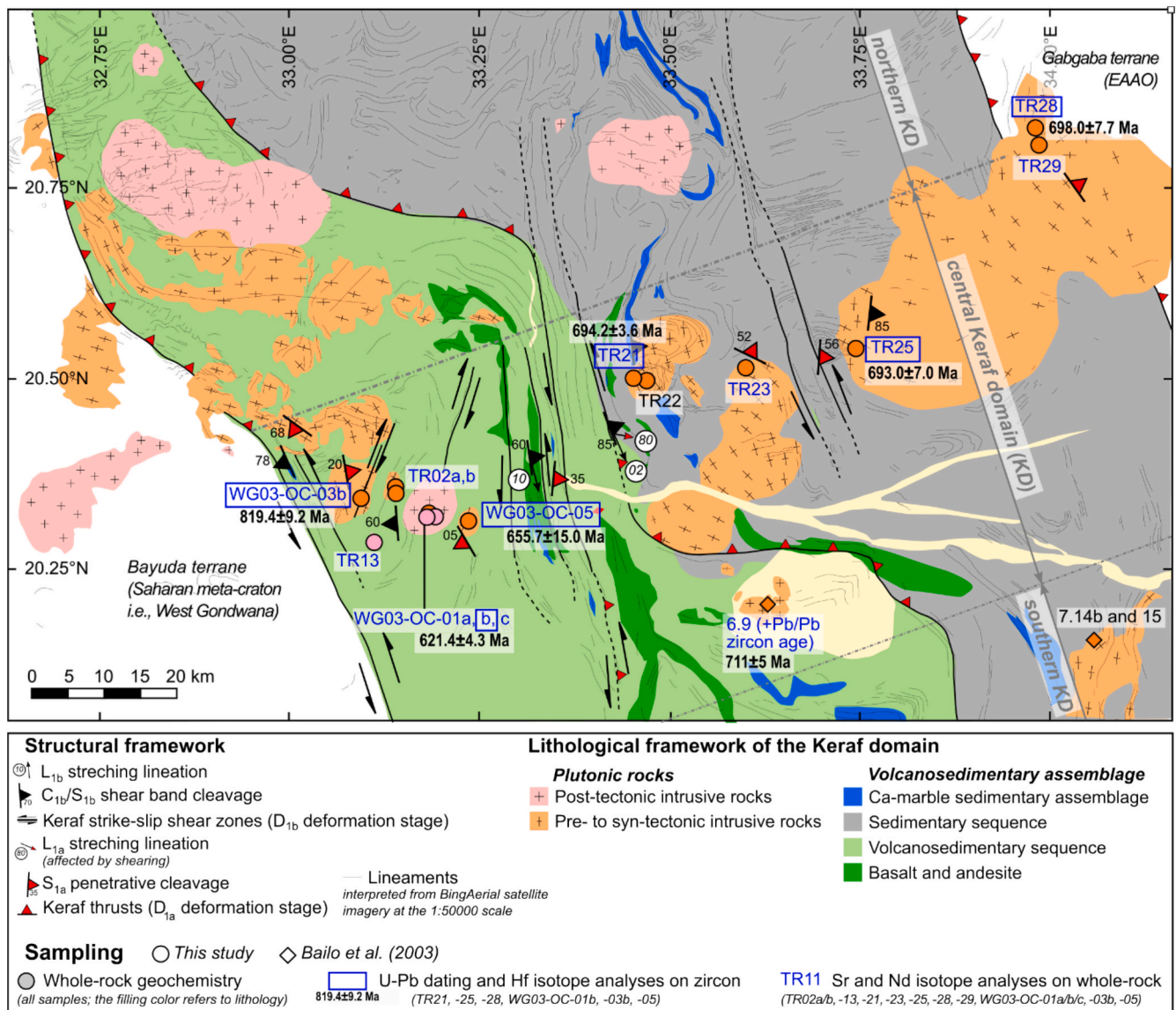


Fig. 2. Geological map of the central Keraf domain, adapted after Bailo et al. (2003), Managem (2011), and field observations (Perret et al., 2021). The extent of the map is reported in Fig. 1. Locations of samples used in this study and corresponding analytical work are indicated (Table 1). Samples from Bailo et al. (2003) located within the mapped area and discussed in the text are also shown. Geographic coordinates are reported in WGS 84.

interpreted as resulting from the closure of an oceanic re-entrant within the Saharan meta-craton north of the Bayuda domain (Fig. 1b; Denkler et al., 1993; Harms et al., 1994; Schandelmeier et al., 1994b; Fritz et al., 2013; El Hur et al., 2017). The coeval evolution of the arc-arc Atmur-Delgo and arc-continent Keraf suture zones led to the development of a complex, regional to micro-scale, L-tectonite-dominated structural framework at the junction between these two structures (Perret et al., 2020, 2022). Support for this interpretation arises from the identification of igneous rocks of island-arc to back-arc geochemical affinities, high-K calc-alkaline to shoshonitic plutonic suites, and amphibolite- to granulite facies metamorphic rocks within the ~ 500 km-long, 50 km-wide, Keraf tectonic assemblage (e.g., Ahmed Suliman, 2000; Bailo et al., 2003; Abdelrahman et al., 2017). Other authors argue, however, that the arc-continent suture zone lies farther west, beyond the Bayuda domain, which is more compatible with the currently accepted western limit of the EAAO (Fig. 1b; Grenholm, 2019 and references therein). Key lines of evidence for such an alternative configuration are (i) the record of Pan-African tectono-metamorphic and magmatic events within the Saharan

meta-craton, west of the currently mapped Keraf suture zone (Fig. 1b; e.g., Ries et al., 1985; Kröner et al., 1987; Rahman et al., 1990; Küster and Liégeois, 2001) and (ii) global plate-tectonic reconstructions of Gondwana (e.g., Vêrard, 2021). These conflicting interpretations therefore raise fundamental questions about the exact position of the arc-continent suture zone and the tectono-magmatic setting of the Keraf domain, which we aim to address in this study.

To this end, we sampled plutonic rocks along an ~ 80 km-long geological transect across the hypothesised arc-continent suture zone within the central Keraf domain (Fig. 2). Building upon the structural model previously established along the same transect (Perret et al., 2021), we combine whole-rock geochemistry, U-Pb geochronology, and Hf-Sr-Nd isotopic analyses to reconstruct the tectono-magmatic evolution of the Keraf domain and evaluate whether it best represents the arc-continent suture zone bounding the ANS or an intra-oceanic arc terrane.

2. Geological setting

2.1. The East African-Antarctic orogeny and greater Gondwana assembly

Jacobs et al. (1998) introduced the concept of the East African-Antarctic orogeny, recognising both temporal and spatial continuity between the East African and Kuunga orogenies (Fig. 1a, b; Kröner, 1993; Stern, 1994; Meert et al., 1995; Jacobs et al., 2003a, 2003b; Kusky et al., 2003; Fritz et al., 2013). The former resulted in the amalgamation of arc terranes within the ANS, the formation of the Mozambique belt further south, and the accretion of continental blocks forming modern parts of Madagascar, Sri Lanka, India, the Seychelles, and East Antarctica to the east, and the Saharan meta-craton, Congo, and Kalahari cratons to the west. The latter is marked by the collision of Australia and another portion of East Antarctica with these earlier assembled blocks.

Most authors define the East African-Antarctic Orogen (EAAO) as the Pan-African (ca. 1000–542 Ma; Kröner, 1984) collisional boundary between West and East Gondwana, leading to assembly of the greater Gondwana (Fig. 1a; Kröner, 1993; Stern, 1994; Jacobs et al., 1998; Kusky et al., 2003; Meert, 2003; Li et al., 2008; Seton et al., 2012; Fritz et al., 2013; Domeier and Torsvik, 2014; Merdith et al., 2017; Grenholm, 2019; Abd El-Wahed and Attia, 2023). However, neither West nor East Gondwana were fully amalgamated before their Cambrian consolidation in their southern parts (Meert, 2003; Gray et al., 2008; Oriolo et al., 2017; da Schmitt et al., 2018; Boniface, 2019), with evidence for detachment and later re-amalgamation of continental blocks occurring after the early Cambrian (Vérard, 2021).

The final suturing of the EAAO, resulting from the closure of the Mozambique Ocean between East and West Gondwana, is thought to have led to the formation of the Nabitah zone, the Baragoi ophiolite klippe, and the Cobuë and Manica belts extending from Saudi Arabia to Kenya and down to Antarctica (Shackleton, 1996). Eastwards, a paleomagnetic study supports the connection between the southeasternmost terrane frontier of the ANS, i.e., the Omani block, and the Indian-Seychelles-Madagascar blocks of East Gondwana as soon as ca. 720 Ma (Antonio et al., 2026), prior to the final amalgamation of the ANS. Westwards, the contact between the EAAO and West Gondwana is defined by a series of arc-continent sutures and ophiolite-decorated belts, notably the Kerat-Kabus-Sekerr-Kinyiki belt stretching from Egypt to Kenya (Fig. 1b; Vearncombe, 1983; Burke and Sengör, 1986; Frisch and Pohl, 1986; Berhe, 1990; Abdelsalam and Dawoud, 1991; Ries et al., 1992; Mosley, 1993; Abdelsalam et al., 1998; Fritz et al., 2013).

2.2. Location of the suture zone between the Arabian-Nubian shield and west Gondwana

The ANS is interpreted as part of a broad Neoproterozoic orogenic system in North Africa and Arabia known as the Saharides (Sengör et al., 2020; Lom et al., 2024). It is composed mainly of juvenile magmatic arcs and ophiolites that are remnants of subduction and obduction processes that consumed the Mozambique Ocean and related ensialic marginal basins (Vail, 1983, 1985; Burke and Sengör, 1986; Almond and Ahmed, 1987; Abdelsalam and Dawoud, 1991; Stern, 1994; Abdelsalam et al., 1995, 1998; Abdelsalam and Stern, 1996; Johnson et al., 2011; Fritz et al., 2013; Abdelrahman et al., 2017). The ANS represents the largest tract of Neoproterozoic juvenile crust worldwide (Johnson, 2014), as shown by extensive U-Pb, Hf, Sr, Nd, Pb and O isotopic datasets (e.g., Andersson et al., 2006; Hargrove et al., 2006; Be'eri-Shlevin et al., 2009a, 2009b; Stern et al., 2010, 2020; Moghazi et al., 2012; Ali et al., 2013; Eyal et al., 2014; Johnson, 2014; Robinson et al., 2014, 2015a, 2015b; Blades et al., 2015; Saeed et al., 2020).

The Kerat domain has commonly been interpreted as a suture zone marking the western boundary between the ANS, i.e., the northern part of the EAAO, and the Saharan meta-craton that belongs to West

Gondwana (Fig. 1b, 2; Stern, 1994; Fritz et al., 2013; Johnson, 2014). The shift from juvenile Hf, Nd, Sr, O and Pb signatures of Neoproterozoic arc terranes of the ANS to the more evolved isotopic compositions of adjacent Archean to Paleoproterozoic cratons is used to delineate this arc-continent suture zone between the ANS and West Gondwana (Ries et al., 1985; Denkler et al., 1993; Sultan et al., 1993; Harms et al., 1994; Sultan et al., 1994; Küster and Liégeois, 2001; Bailo et al., 2003; Küster et al., 2008; Evuk, 2013; Evuk et al., 2014; Johnson, 2014; Evuk et al., 2017).

Nevertheless, Pan-African tectonic, metamorphic, and magmatic events are recorded west of the Kerat domain, within the Bayuda and Sabaloka domains, which are both parts of the Saharan meta-craton (Fig. 1b; e.g., Ries et al., 1985; Kröner et al., 1987; Rahman et al., 1990; Küster and Liégeois, 2001). Evidence for such events challenges the interpretation of the Kerat domain as the primary suture zone and indicates that much remains to be resolved regarding the tectono-magmatic evolution of the eastern margin of West Gondwana (e.g., Abdelsalam and Stern, 1996; Ahmed Suliman, 2000; Küster and Liégeois, 2001; Evuk et al., 2014).

2.3. Lithological-tectonic evolution of the Kerat domain

The stratigraphic and plutonic record of the central Kerat domain has been interpreted primarily as reflecting magmatic arc and/or back-arc basin development, represented by volcanoclastic sequences, andesitic flows, and granitoid plutons (e.g., Schandelmeier et al., 1994b; Stern, 1994; Abdelsalam et al., 1995; Abdelrahman et al., 2017; Perret et al., 2021), and, to a lesser extent, passive continental margin deposition, expressed by carbonate-rich and siliciclastic metasedimentary rocks (e.g., Stern et al., 1993). Initial convergence and ocean closure occurred along a north-striking magmatic arc active between ca. 800–750 Ma (Abdelsalam et al., 1998; Bailo et al., 2003). Subsequent oblique convergence between ca. 750–640 Ma resulted in arc-continent collision at the scale of the whole domain (Abdelsalam et al., 1998). This tectonic event is assumed to be responsible for west-verging thrusting and the development of a regional-scale, penetrative planar fabric within the central Kerat domain (Fig. 2). Progressive deformation subsequently evolved from thrusting into sinistral shearing between ca. 640–580 Ma, producing major strike-slip shear zones throughout the region (Fig. 2; Abdelsalam et al., 1998; Perret et al., 2021). Undeformed granitoids intruded the deformed and metamorphosed lithologies across the central Kerat domain during the late to post-collisional period (ca. post-620 Ma). These bodies occur either as lenses between shear zones or as crosscutting plutons (Fig. 2; Abdelsalam et al., 1998; Perret et al., 2021).

3. Sampling and analytical techniques

3.1. Field observations and sampling

We present geochemical and isotopic data from fourteen plutonic rocks exposed along a west-southwest-trending transect across the central Kerat domain (Fig. 2). Although mafic metavolcanic units were also sampled during this investigation, they were significantly weathered and/or altered and contained no recoverable zircon. They were therefore excluded from the final dataset. Field data were collected using an early version of the GEOL-QMAPS digital geological mapping solution (Perret et al., 2024; Perret, 2025). Sample locations and analytical methods applied to each specimen are listed in Table 1.

3.2. Analytical techniques

3.2.1. Whole-rock geochemistry

Whole-rock major and trace element compositions of the fourteen samples were determined by inductively coupled plasma optical emission spectrometry (ICP-OES) and mass spectrometry (ICP-MS), using Thermo Fisher iCAP 6500 and Thermo Fisher iCAP Q spectrometers,

Table 1

List of studied samples. Geographic coordinates are reported in WGS 84. The analytical work carried out for each sample is indicated.

Sample ID	Easting WGS84	Northing WGS84	Rock type*	Results presented in this study		
				Whole-rock geochemistry (Figs. 5-6; Supplementary Data, Table S1)	U-Pb and Hf isotopes on zircon (Figs. 7-8; Supplementary Data, Tables S3-4)	Sr and Nd isotopes on whole-rock (Fig. 8; Supplementary Data, Table S7)
Group 1 (subgroup 1.1)						
TR02a	33.1402	20.3577	Quartz-diorite	x		x
WG03-OC-01c	33.1837	20.3234	Monzodiorite	x		x
WG03-OC-03b	33.0950	20.3431	Quartz-diorite	x	x	
Group 1 (subgroup 1.2)						
TR02b	33.1405	20.3578	Granodiorite	x		x
TR21	33.4512	20.5002	Quartz-monzodiorite	x	x	x
TR22	33.4698	20.4969	Quartz-monzodiorite	x		
TR23	33.6000	20.5142	Granodiorite	x		x
TR25	33.7448	20.5399	Diorite	x	x	x
TR28	33.9806	20.8292	Quartz-monzodiorite	x	x	x
TR29	33.9851	20.8068	Granodiorite	x		x
WG03-OC-05	33.2364	20.3132	Granodiorite	x	x	x
Group 2						
TR13	33.1125	20.2853	Monzogranite	x		x
WG03-OC-01a	33.1838	20.3233	Granodiorite	x		x
WG03-OC-01b	33.1838	20.3233	Monzogranite	x	x	x

* Protolith lithologies are mentioned for ortho-derived rocks which have undergone deformation and metamorphism, relying on [Streckeisen \(1976\)](#) classification.

respectively, at the Service d'Analyse des Roches et des Minéraux (SARM, CRPG-CNRS, Vandœuvre-lès-Nancy, France; [Table 1](#)). Sample preparation procedures, analytical conditions, and detection limits are described in [Carignan et al. \(2001\)](#). Relative analytical uncertainties are approximately 2 % for major elements and between 5 % and 20 % for trace elements, depending on concentration. Whole-rock S concentrations were measured using a CS-analyser with an approximate limit of detection of 0.01 wt%. Whole-rock Au concentrations for three samples were determined by ALS via fire assay and ICP-AES, with a detection limit of 1 ppb. Results are provided in [Supplementary Data \(Table S1\)](#).

3.2.2. Zircon separation and imagery

Six samples were selected for U–Pb dating and Hf isotope analyses on zircon ([Table 1](#)). Zircon separation was performed at the Centre de Recherches Pétrographiques et Géochimiques (CRPG-CNRS, Vandœuvre-lès-Nancy, France). Samples were crushed successively with jaw and roll crushers. Zircon grains were extracted from the 75–125 µm and 125–250 µm size fractions using a Frantz magnetic separator and diiodomethane (density = 3.32 g/cm³) for heavy-liquid separation. Representative zircon grains of variable morphology were hand-picked in ethanol under a binocular microscope, mounted in epoxy resin and polished to expose their cores. Back-scattered electron (BSE) and cathodoluminescence (CL) images were obtained using a TESCAN VEGA3 LM conventional SEM coupled with a GATAN ChromaCL2 system (10 kV acceleration voltage, 5nA current intensity) at the GeoRessources research centre (Vandœuvre-lès-Nancy, France). Following BSE and CL imaging, grains representative of the different zircon populations identified in each sample, based on morphology and internal textures, were selected for U–Pb–Hf dating.

3.2.3. U–Pb dating and Hf isotope analyses of zircon

For the six selected samples, zircon grains were first analysed for U–Pb geochronology and subsequently for Hf isotopes by laser ablation-inductively coupled plasma-sector field mass spectrometry (LA-ICP-SF-MS). Spot locations were selected from CL images to target texturally

homogeneous domains and avoid mixed core-rim analyses wherever possible.

A first U–Pb analytical session was performed with a Resonetics M50 193 nm Excimer laser system coupled to a Thermo-Scientific ELEMENT 2 LA-SF-ICP-MS at Goethe University (Frankfurt, Germany) for samples WG03-OC-01a, -03b and -05. A second U–Pb analytical session used a Thermo-Scientific Element XR LA-SF-ICP-MS coupled to a 193 nm ArF Excimer laser (Analyte Exite +, Teledyne Photon Machines) at Karlsruhe Institute of Technology (Karlsruhe, Germany) for samples TR21, TR25 and TR28.

Hf isotopes were analysed with a Resonetics M50 193 nm Excimer laser system coupled to a Thermo-Scientific NEPTUNE LA-SF-ICP-MS at Goethe University (Frankfurt, Germany) for all samples. Instruments and processing parameters are described by [Gerdes and Zeh \(2009\)](#) with modifications in [Zeh and Gerdes \(2012\)](#). Laser spots of 40–50 µm diameter for Hf isotope analyses were placed over U–Pb laser spots of 25–33 µm-diameter. Zircon standards used in the bracketing procedure are: (i) GJ-1 ([Jackson et al., 2004](#)), Plešovice ([Sláma et al., 2008](#)), BB ([Santos et al., 2017](#)) and Kaap Valley ([Takenaka et al., 2015](#)) for U–Pb dating; (ii) GJ-1 ([Morel et al., 2008](#)) and Temora 1 and 2 ([Woodhead et al., 2004](#); [Xu et al., 2004](#); [Woodhead and Hergt, 2005](#); [Wu et al., 2006](#)) for Hf isotopes. $\epsilon_{\text{Hf}(t)}$ values were calculated back to ²⁰⁶Pb/²³⁸U single zircon concordant dates for magmatic zircon grains.

U–Pb and Hf isotope laser-spot positions superimposed to zircon textures on CL images are presented in [Supplementary Data](#) (Photograph Plate S2). Analytical results and raw data-processing procedures are available in [Supplementary Data](#) (Tables S3 and S4), as well as reference zircon standard data (Tables S5 and S6).

3.2.4. Sr and Nd whole-rock isotope analyses

Twelve samples were investigated for their whole-rock Sr and Nd isotopic compositions ([Table 1](#)). Analyses were carried out at the CRPG-CNRS (Vandœuvre-lès-Nancy, France). Between 100–200 mg of bulk powders were dissolved by adding 4 mL of concentrated HNO₃ and 1 mL of concentrated HF and heating for one to two days at 115 °C. Samples

were then dried down and heated again at 125 °C for one day after the addition of HCl. They were then mixed with 2 mL HNO₃. Chromatographic columns were then used to separate Sr, Rb and rare earth elements (REE) following the methods of Pin et al. (1994) and Pin and Zalduegui (1997) which involve the TRU.Spec resin to extract the REE from the matrix, the Sr.Spec to separate Sr and the Eichrom LN.Spec to isolate Nd and Sm.

For Sr isotopic compositions, the resulting solutions were analysed with a Triton Plus thermal ionisation mass spectrometer. Of the five Faraday cages used, one monitored Rb. Internal normalisation was applied using $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$, to correct for instrumental fractionation. Standard NBS987, certified by the National Institute of Standards and Technology, was analysed after every third sample for quality assurance. Sr total chemistry blanks consistently yielded values below 100 pg which were negligible relative to the quantities of Sr measured in the samples.

For Nd isotopic compositions, the solutions were analysed with a Neptune multi-collection ICP-MS. Internal normalisation was applied using $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, in order to correct for instrumental fractionation (Luais et al., 1997). Standard JNdi (Tanaka et al., 2000) was analysed after every third sample for quality assurance. Nd and Sm total chemistry blanks were highly reproducible, consistently yielding values below 100 pg which were negligible relative to the quantities of Nd and Sm measured in the samples.

Initial ($^{87}\text{Sr}/^{86}\text{Sr}$)_i, ($^{143}\text{Nd}/^{144}\text{Nd}$)_i and $\varepsilon\text{Nd}_{(t)}$ values were calculated by back-projection to zircon U-Pb crystallisation ages. For undated samples, crystallisation ages were inferred from structural and geochemical similarity to dated samples, following the definitions of Group 1 (pre- to syn-tectonic plutonic rocks, subdivided into 1.1 and 1.2 based on geochemical distinction) and Group 2 (post-tectonic intrusions; see section 4). Results are presented in Supplementary Data (Table S7).

4. Results

In this section, we introduce two groups to classify plutonic rocks sampled across the central Kerf domain based on their relative timing with respect to the deformation stages recognised in the area (see subsection 4.1; Perret et al., 2021). We further distinguish two subsets of pre- to syn-tectonic magmatic rocks relying on their geochemical compositions (see subsection 4.2) and present their respective U-Hf-Sr-Nd isotopic signatures (see subsections 4.3, 4.4).

4.1. Field description of Group 1 and Group 2 plutonic rocks

4.1.1. Group 1

Schistose, low-grade basalt and andesite interlayered with volcaniclastic and sedimentary rocks, are the dominant lithology in the studied area (Figs. 2, 3a). They are strongly deformed and completely transposed by the regional schistosity related to collision-related thrusting (S_{1a}; Figs. 2, 3a).

Group 1 plutonic rocks, dioritic to granitic in composition, intrude the volcanosedimentary series (Fig. 2) and are variably deformed, locally showing a gneissic foliation that affects the outcrop unevenly (Fig. 3c, d). A down-dip stretching lineation is locally preserved where the gneissic foliation is best expressed (Fig. 3d). Later Kerf strike-slip shear zones, visible at the regional scale, may cut across these plutons, as expressed by the penetrative fabric at the macroscopic scale (Figs. 2, 3b).

4.1.2. Group 2

Group 2 consists of granitoid intrusive bodies of variable size in map view (Fig. 2) that typically crop out as up to tens-of-metres-high autochthonous blockfields forming a granitic chaos (Fig. 3e). These rocks are undeformed: except for a weak magmatic fabric that is rarely observed, they show no anisotropic feature at the macroscopic scale. Feldspar phenocrysts are present in a quartz-(feldspar)-rich term

(WG03-OC-01a), which shows diffuse contacts with the quartz-enriched, coarse-grained host granitoid (WG03-OC-01b; Fig. 3f). Quartz-poor xenoliths (WG03-OC-01c) are also enclosed in the intrusion (Fig. 3f).

4.2. Whole-rock geochemistry

4.2.1. Group 1

Based on major and minor element signatures, Group 1 plutonic rocks are split into two subgroups.

4.2.1.1. Subgroup 1.1. With SiO₂ contents ranging from 49.5 to 55.7 wt % and Mg# between 39.0–46.4 %, Subgroup 1.1 plutonic rocks are mafic to slightly intermediate (Supplementary Data, Table S1). In terms of major elements, they classify as quartz-diorite (Fig. 3a, b; Supplementary Data, Table S1). In the feldspar modal classification ternary plot, they appear to be weakly potassic as they plot along the plagioclase feldspar trend (Fig. 4a). Subgroup 1.1 granitoids are met-aluminous (Fig. 4b).

Chondrite-normalised REE patterns show flat light REE (LREE) segments ($\text{La}_\text{N}/\text{Nd}_\text{N}$ in the range of ~ 1.0 – 1.3) and even flatter high REE (HREE) segments ($\text{Dy}_\text{N}/\text{Lu}_\text{N}$ in the range of ~ 1.1 – 1.3 ; Fig. 5a, d). Sr and Eu anomalies are defined by $\text{Eu}_\text{N}/\text{Eu}_\text{N}^* = \text{Eu}_\text{N}/(\text{Sm}_\text{N} \cdot \text{Gd}_\text{N})^{0.5}$ and $\text{Sr}_\text{N}/\text{Sr}_\text{N}^* = \text{Sr}_\text{N}/(\text{Pr}_\text{N} \cdot \text{Nd}_\text{N})^{0.5}$ ratios, respectively. Subgroup 1.1 samples may present relatively high Sr (e.g., ~ 3) and moderately positive Eu anomalies (e.g., ~ 1.2 for the quartz-diorite sample; Fig. 5e). Normalised to primitive mantle, their compositions in incompatible trace elements display a low to intermediate enrichment in the most incompatible ones (i.e., left-hand side of the diagram, mostly the large-ion lithophile elements, or LILE: Cs, Rb, Ba, K, Pb, Sr), that vary from one sample to another, whereas the more compatible ones (i.e., right-hand side of the diagram, mostly the high-field strength elements, or HFSE: Th, U, Nb, Ta, Zr, Hf, Ti, Y) show similar patterns for all samples, with an enrichment factor of ~ 7 – 10 (Fig. 5f). These samples display consistently a low, negative Ti anomaly. Immobile element Th/Nb and La/Nb ratios range between 0.06–0.95 and 2.8–3.9, respectively (Supplementary Data, Table S1). The mafic monzodiorite xenolith enclosed in a post-tectonic monzogranite pluton has broadly the same major and trace element composition as Subgroup 1.1 granitoids (Figs. 3f, 4, 5; Supplementary Data, Table S1).

4.2.1.2. Subgroup 1.2. Major element compositions classify Subgroup 1.2 granitoids as an intermediate to felsic, metaluminous, I-type diorite-granodiorite-granite suite (Figs. 3b–d, 4), with Mg# values ranging from 15.1 to 30.4 %, except for one outlier at 7.6 % (Supplementary Data, Table S1). They consistently display chondrite-normalised REE patterns with a steeper decreasing slope for LREE than for HREE ($\text{La}_\text{N}/\text{Nd}_\text{N}$ in the range of ~ 1.5 – 3.3 , $\text{Dy}_\text{N}/\text{Lu}_\text{N}$ in the range of ~ 0.8 – 1.8 ; Fig. 5b, d). The steeper the LREE segment, the higher the LREE enrichment relative to chondrite (e.g., $\text{La}_\text{N}/\text{Nd}_\text{N} = 3.3$ and La enrichment factor of 200 for WG03-OC-05 granodiorite vs $\text{La}_\text{N}/\text{Nd}_\text{N} = 1.8$ and enrichment factor of 75 for TR29 granodiorite; Fig. 5b, e). Moreover, half of the studied granitoids have a Sr anomaly between 1.5 and 2, with Eu anomaly ranging between 0.7 and 1.2 (Fig. 5e). Finally, they tend to display more evolved trace element patterns normalised to the primitive mantle than Subgroup 1.1 granitoids, which match the signatures of Bailo et al. (2003)'s samples in the southern Kerf domain (Fig. 5g). Overall, enrichment in LILE is stronger than for Subgroup 1.1, and HFSE patterns evolve from Subgroup 1.1-like, ~ 7 – 10 enrichment factor to primitive mantle-like. With respect to Subgroup 1.1, these samples display a stronger negative Ti anomaly, due to stability of Fe-Ti oxides, and Zr-Hf relative enrichment, due to the greater abundance of zircon (Fig. 5g). Immobile element Th/Nb and La/Nb ratios are higher than for Subgroup 1.1, and range between 0.80–2.5 and 4.0–7.5, respectively, with WG03-OC-05 being a clear outlier (ratio values of 5.0 and 18.4, Supplementary Data, Table S1).

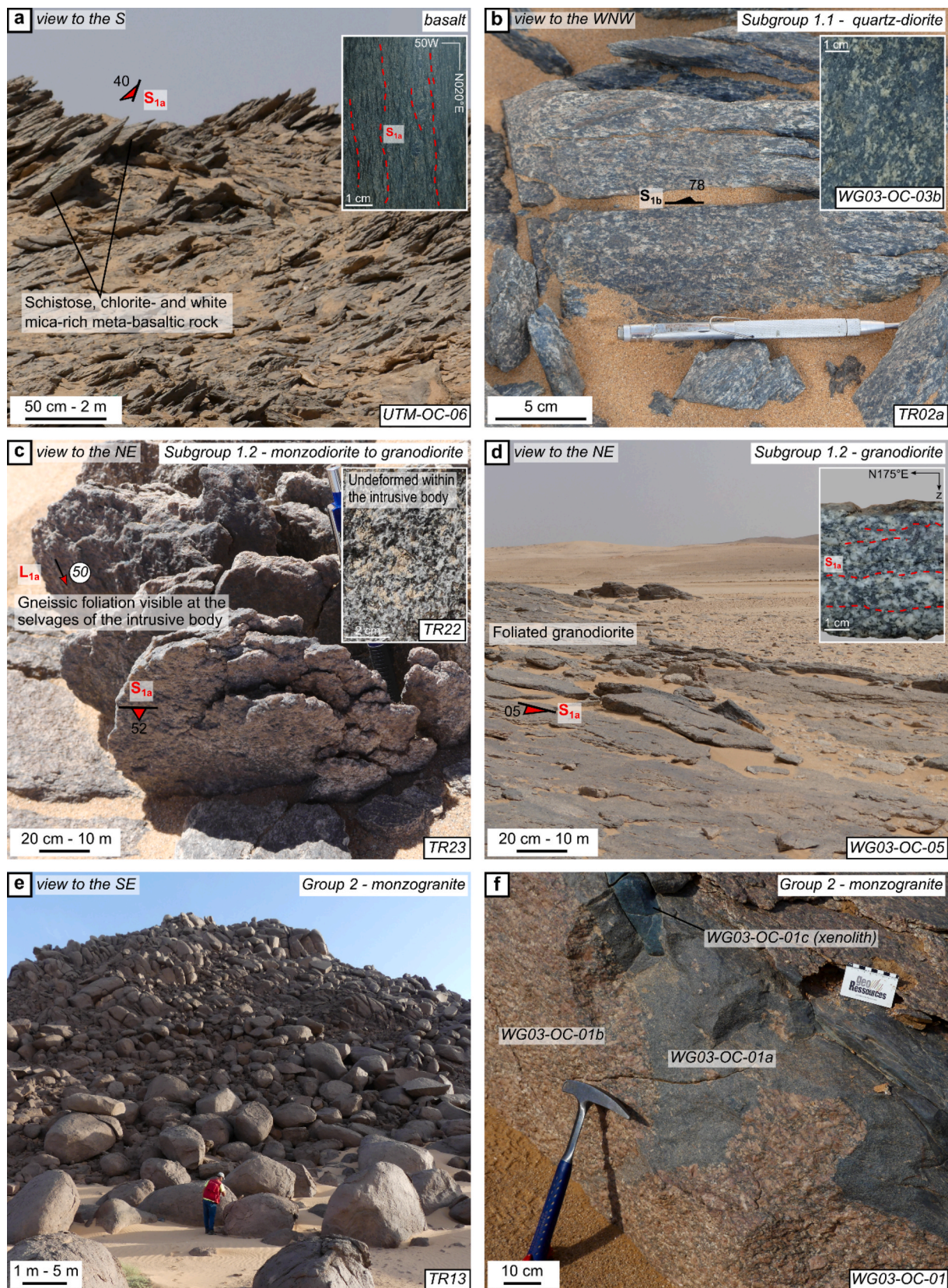


Fig. 3. Pre- to syn-tectonic (Group 1) and post-tectonic (Group 2) plutonic rocks of the central Keraf domain. Outcrop location refers to sample IDs in Fig. 2. Structural indices refer to regional deformation stages (Perret et al., 2021). **a.** Schistose basalt, affected by the regional thrusting-related penetrative cleavage (S_{1a}) and intruded by the various plutonic bodies described below. **b.** Granitoid affected by shear-band cleavage (S_{1b}) along a late collisional shear zone. **c-d.** Granitoids foliated along thrusting-related, regional penetrative fabric. **e.** Undeformed granitoids cutting across the regional structures. **f.** The latter display a mingling texture between an intermediate and a more felsic terms (WG03-OC-01a and b, respectively), and enclose monzodiorite xenoliths (WG03-OC-01c).

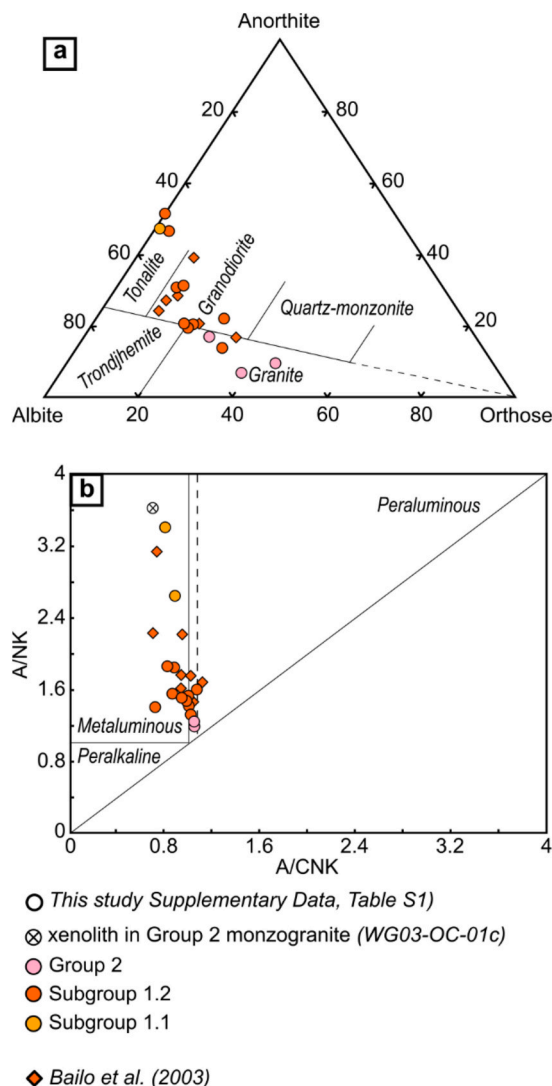


Fig. 4. Whole-rock major element compositions of samples from this study (circles) and Keraf arc-related granitoids from Bailo et al. (2003) (diamonds). **a.** Feldspar modal classification for silicic plutonic rocks with > 10 % modal quartz (O'Connor, 1965). **b.** A/CNK vs A/NK crossplot for all samples (Shand, 1943). The dashed line indicates the uncertainty field between metaluminous and peraluminous domains. Whole-rock data are compiled in Supplementary Data (Table S1).

4.2.2. Group 2

Group 2 intrusions are felsic (> 62 wt% SiO₂), with relatively low Mg# (5.0 to 14.7 %) (Supplementary Data, Table S1) and plot in the granite field of the feldspar modal classification ternary plot (Fig. 4a). They mostly consist of coarse-grained monzogranite to granodiorite (WG03-OC-01b and -01a, respectively; Fig. 3f; Table 1). These rocks are slightly peraluminous (Fig. 4b). They show different chondrite-normalised REE patterns (Fig. 5c): TR13 monzogranite displays both relatively steep LREE and HREE segments (La_N/Nd_N = 2.8 and Dy_N/Lu_N = 1.8; Fig. 5d), whereas WG03-OC-01b monzogranite shows a bell-shaped concave-down pattern with a decreasing LREE slope and a moderate, positive HREE slope (La_N/Nd_N = 1.5 and Dy_N/Lu_N = 0.6; Fig. 5d). The WG03-OC-01a granodiorite REE pattern is similar to TR13 for LREE, but the HREE segment is almost flat (La_N/Nd_N = 2.5 and Dy_N/Lu_N = 1.1; Fig. 5d). Regardless, Group 2 intrusions consistently have no to negative Sr and Eu anomalies (~0.3–1 and ~0.65–0.75, respectively; Fig. 5e). These late plutons display a similar pattern to Subgroup 1.2

granitoids, with a slightly higher enrichment in LILE (Fig. 5h). Note (i) the strong Th-U enrichment for WG03-OC-01b sample, (ii) HFSE concentrations spanning over the entire range covered by Subgroup 1.1 and 1.2 granitoids, and (iii) a negative Ti anomaly similar to that reported in Subgroup 1.2 (Fig. 5h). For the three samples considered, Th/Nb ratios are 0.89, 2.8 and 17.1, and La/Nb ratios are 4.3, 3.0 and 1.1 (Supplementary Data, Table S1).

4.3. U-Pb ages and Hf isotope analyses of zircon

We followed zircon morphology, textural and zoning information to classify zircon types and distinguish between populations and interpret their geological origin (Corfu, 2003; Zheng et al., 2022). We use the nomenclature of Miller et al. (2007) to describe the inherited (i.e., that 'survive(d) anatexis events'), xenocrystic (i.e., 'incorporated from the surrounding host rock during transit and emplacement'), antecrystic (i.e., 'crystallized from an earlier pulse of magma and which are incorporated in a later pulse(s)'), or autocrystic (i.e., grown 'only within the youngest intrusive pulse') nature of dated magmatic zircon grains. U concentration and Th/U ratio thresholds attesting the magmatic nature of zircon domains analysed are based on the compilation by Kirkland et al. (2015). Fig. 6 provides cathodoluminescence images of representative grains extracted from Subgroups 1.1 and 1.2 and Group 2 intrusions, highlighting key textural and zoning patterns. The geochemical distinction between Subgroups 1.1 and 1.2 aligns with different zircon patterns and crystallisation ages for dated plutons.

Weighted mean ²⁰⁶Pb/²³⁸U dates were calculated using IsoplotR (Vermeesch, 2018), based on the original Isoplot program (Ludwig, 2007). Analyses outside the 95–105 % concordance range, inherited zircon grains, analyses affected by Pb loss or later tectono-metamorphic resetting, and analyses with Th/U < 0.2 were excluded. Only analyses interpreted to belong to the autocrystic population were included in each weighted mean calculation (Supplementary Data, Table S3).

Weighted mean ²⁰⁶Pb/²³⁸U dates were preferred over Concordia U-Pb dates as several samples contain zircon populations that define arrays of concordant analyses distributed along the Concordia curve rather than tightly clustered age populations. In such cases, a single Concordia date can obscure geologically meaningful age dispersion related to prolonged zircon crystallisation, antecryst recycling, and magma chamber evolution. Weighted mean date plots allow identification of discrete zircon growth populations and, in particular, the youngest coherent autocrystic population interpreted to record the final crystallisation stage of the intrusion. For comparison, Concordia ages calculated from the same selection of analyses are reported alongside weighted mean ages in what follows, and Wetherill Concordia plots for all samples are provided in Supplementary Data (Table S3). Obtained ages are consistent between the two approaches.

4.3.1. Subgroup 1.1

4.3.1.1. WG03-OC-03b quartz-diorite. Recovered zircon grains form a single population. They are subhedral to euhedral, sometimes exceeding 250 μm long, with length-to-width ratios of 1:1 to 1:3. Zircon displays mostly igneous sector zoning, with oscillatory growth bands, both typical of igneous zircon. Zoning is generally regular from core to rim, consistent with autocrystic growth, although the presence of antecrystic components cannot be ruled out. Oscillatory zoning can indeed result from continuous or episodic crystal growth under similar chemical conditions during either a single magmatic event or multiple magmatic pulses. In the former case, autocrystic zircon can form, whereas in the latter, antecrystic cores with concordant younger autocrystic overgrowths may develop, and textural patterns alone are insufficient to distinguish between these possibilities. Some grains contain rare voids and mineral inclusions that reflect localised post-crystallisation alteration. These domains were avoided during analytical spot selection.

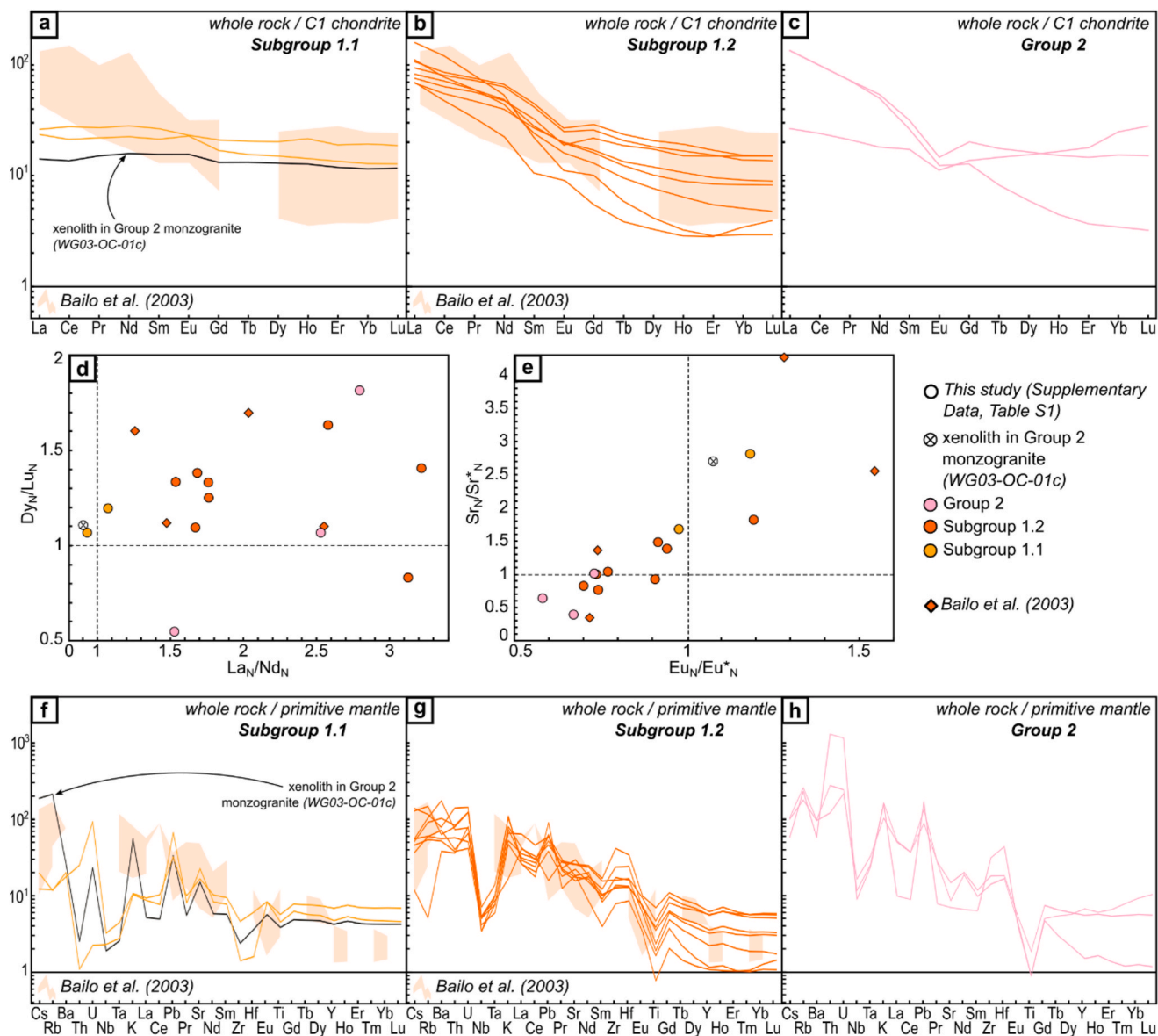


Fig. 5. Whole-rock trace element compositions of samples from this study and Kerf arc-related granitoids from Bailo et al. (2003). Whole-rock data are compiled in [Supplementary Data \(Table S1\)](#). **a–c.** C1 chondrite-normalised rare earth trace element diagrams, with normalisation values from Sun and McDonough (1989), for **(a)** Subgroup 1.1, **(b)** Subgroup 1.2 and **(c)** Group 2 samples. **d.** C1 chondrite-normalised La_N/Nd_N vs Dy_N/Lu_N ratio crossplot (Sun and McDonough, 1989). The dashed lines denote flat LREE and HREE patterns. **e.** Eu_N/Eu*_N vs Sr_N/Sr*_N anomaly crossplot. Eu and Sr anomalies are calculated as the ratio between the C1 chondrite-normalised Eu_N and Sr_N values with Eu*_N = (Sm_N,Gd_N)^{0.5} and Sr*_N = (Pr_N,Nd_N)^{0.5} respectively (Sun and McDonough, 1989). The dashed lines mark null Eu and Sr anomalies. **f–h.** Primitive mantle-normalised extended spider diagrams for incompatible trace elements, with normalisation values from Sun and McDonough (1989), for **(f)** Subgroup 1.1, **(g)** Subgroup 1.2 and **(h)** Group 2 granitoids.



Fig. 6. Representative cathodoluminescence images of dated zircon grains.

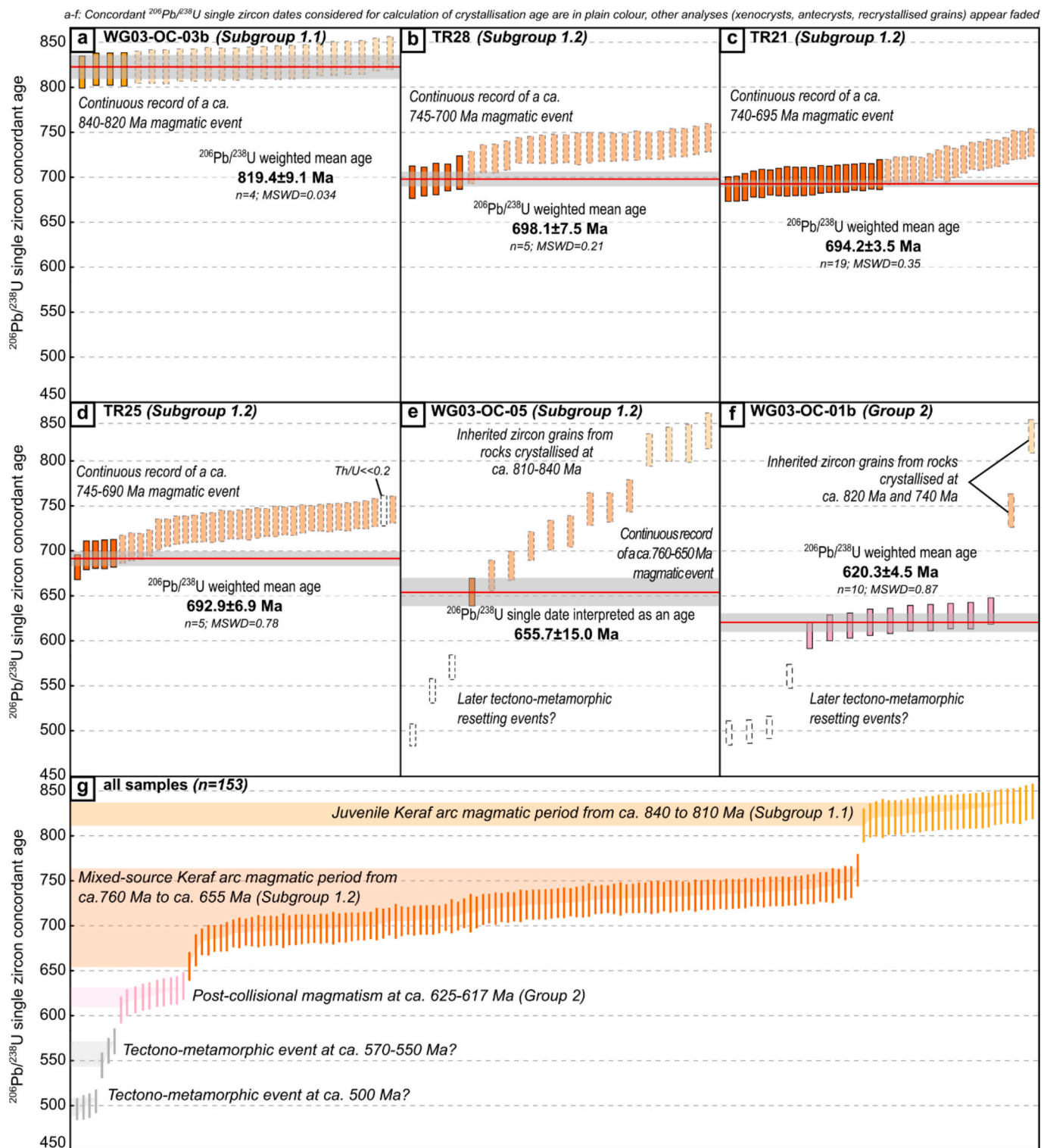


Fig. 7. Weighted mean plots (2σ error) of $^{206}\text{Pb}/^{238}\text{U}$ single zircon ages for dated samples. Data outside the 95–105 % concordance range were excluded. Plots are ordered by decreasing crystallisation age and colour-coded by zircon generation. Bars with solid-line outlines were used in the $^{206}\text{Pb}/^{238}\text{U}$ weighted mean crystallisation age calculations. **a.** WG03-OC-03. **b.** TR28. **c.** TR21. **d.** TR25. **e.** WG03-OC-05. **f.** WG03-OC-01b. **g.** all samples combined. Plots have been generated using IsoplotR (Vermeesch, 2018), based on the original Isoplot program (Ludwig, 2007). Cathodoluminescence and BSE-SEM images showing zircon morphology, internal zoning, and data spot positions are provided in Supplementary Data (Photograph Plate S2). U-Pb data are compiled in Supplementary Data (Table S3).

Twenty-five U-Pb measurements were made on twenty-five zircon grains, avoiding any altered domains (Supplementary Data, Table S3). U contents are low (up to 131 ppm), and Th/U ratios are magmatic, varying between 0.4 and 0.7. $^{206}\text{Pb}/^{238}\text{U}$ single zircon dates span continuously over ca. 840–820 Ma (Fig. 7a). This spread is interpreted in

terms of magmatic processes as there is neither textural, geochemical, nor analytical evidence for the existence of any tectono-metamorphic resetting or analytical disturbance which could account for such a Concordia-parallel Pb loss. We interpret a $^{206}\text{Pb}/^{238}\text{U}$ weighted mean age of 819.4 ± 9.1 Ma from the well-clustered four youngest concordant

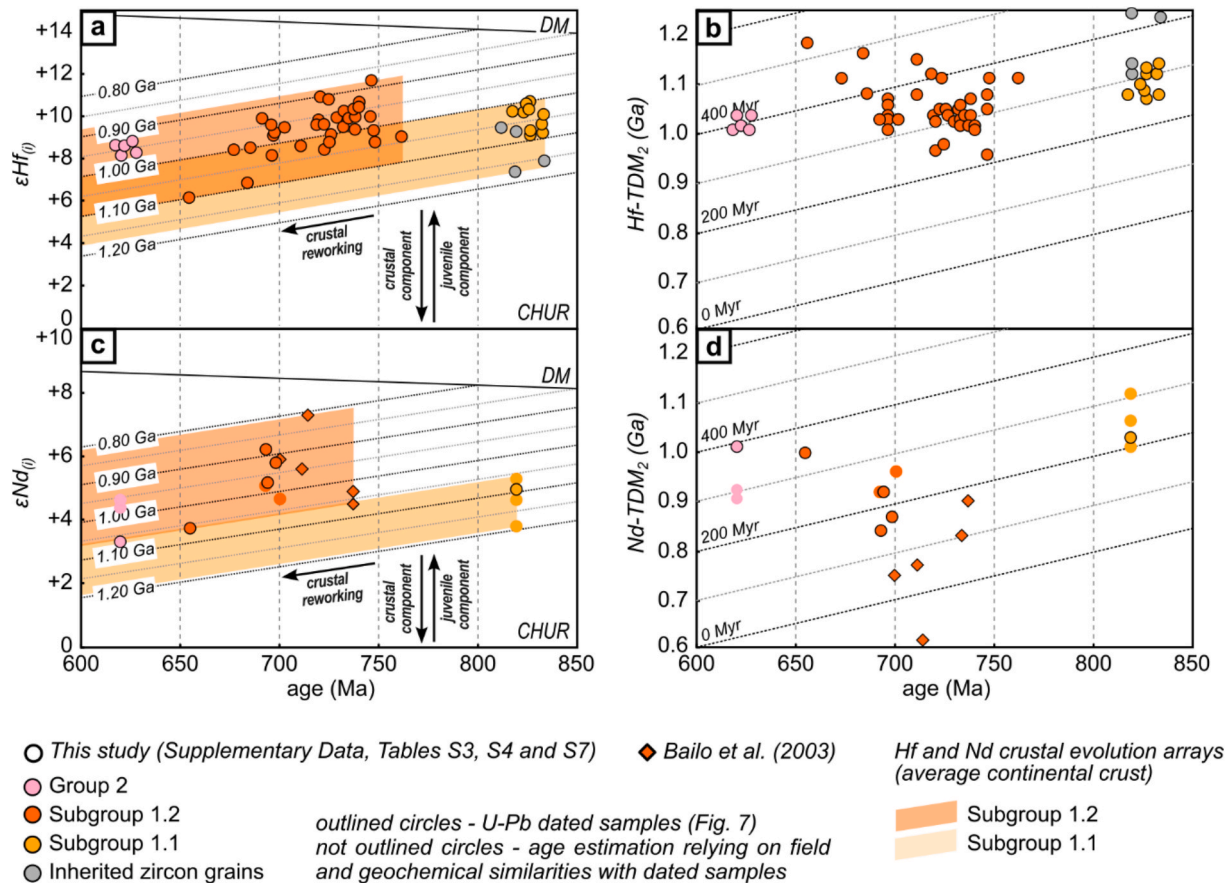


Fig. 8. Crossplots of Hf and Nd isotopic data from this study (circles) and Keraf arc-related granitoids from Bailo et al. (2003) (diamonds). Plots show (a) $^{238}\text{U}/^{206}\text{Pb}$ single zircon age vs their initial zircon epsilon hafnium composition $\epsilon\text{Hf}_{(t)}$ (depleted mantle DM evolution from Blichert-Toft and Puchtel, 2010), (b) crystallisation age vs Hf- $\text{T}_{\text{DM}2}$ model-age, (c) crystallisation age vs initial whole-rock epsilon neodymium composition $\epsilon\text{Nd}_{(t)}$ (depleted mantle DM evolution from Goldstein et al., 1984), and (d) crystallisation age vs Nd- $\text{T}_{\text{DM}2}$ model-age. Crystallisation ages in (b–d) correspond to U-Pb weighted mean ages from this study or Pb/Pb zircon ages (Bailo et al., 2003). (a) and (c): CHUR denotes chondritic uniform reservoir in panels. Reference $\text{T}_{\text{DM}2}$ age evolution lines for the average continental crust (i.e., isotopic trajectories of crustal reworking) were calculated using $^{176}\text{Lu}/^{177}\text{Hf} = 0.0113$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.120$ (Liew and Hofmann, 1988; Blichert-Toft and Puchtel, 2010) between 0.80 and 1.20 Ga with a 0.05 Gyr step, respectively. (b) and (d): Expected trends for juvenile mantle-derived input, crustal contamination (i.e., reworking of previous juvenile crust) and crustal reworking are shown. Lines indicative of reference crustal residence time between 0 and 500 Myr are shown in. U-Pb, Hf and Nd isotopic data are compiled in Supplementary Data (Tables S3, S4 and S7, respectively).

$^{206}\text{Pb}/^{238}\text{U}$ single zircon dates (MSWD = 0.034; Fig. 7a) as the crystallisation age. The same analyses yield a concordant age of 816.6 ± 6.2 Ma (MSWD = 0.57; Supplementary Data, Table S3). Older, ca. 840–820 Ma single zircon dates likely relate to antecrysts recording cascading magmatic pulses, or continuous magmatism, over this ~ 20 Myr-long magmatic period.

Ten Hf isotope analyses on individual zircon grains with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 833–817 Ma yield $\epsilon\text{Hf}_{(t)}$ values ranging from $+9.1 \pm 0.8$ to $+10.7 \pm 0.8$ (mean value of $+9.9 \pm 0.8$; Fig. 8a; Supplementary Data, Table S4). Two-stage model ages $\text{T}_{\text{DM}2}$ span 1.12–1.05 Ga (Fig. 8b).

4.3.2. Subgroup 1.2

4.3.2.1. TR28 quartz-monzodiorite. Recovered zircon grains form a single population. They are subhedral, up to 200 μm long with a length-to-width ratio of 1:1.5 to 1:4. Many grains enclose numerous voids and mineral inclusions, suggesting subsequent hydrothermal alteration. Unaltered grains and pristine domains preserve igneous sector and growth zoning. No distinct xenocryst cores are identified texturally. The antecryst or autocryst

nature of these zircon grains remains unclear from textural evidence only. Fractured and altered grains show distorted internal textures and patchy zoning overprinting oscillatory zoning; these grains were excluded from U-Pb and Lu-Hf analyses.

Twenty-nine U-Pb measurements were made on twenty-nine grains (Supplementary Data, Table S3). Two discordant datapoints (U142, U170) were disregarded. U concentrations are < 250 ppm and Th/U ratios ~ 0.4–0.9, compatible with a magmatic origin, in agreement with zoning patterns observed. Concordant $^{206}\text{Pb}/^{238}\text{U}$ single zircon dates range from 743 ± 16 Ma to 694 ± 19 Ma with significant overlapping arrays. These data are interpreted to reflect recycling of zircon antecrysts incorporated in later magmatic pulses. Because only one analysis was made per grain, the presence of younger autocrystic overgrowths cannot be confirmed. The five youngest concordant single zircon dates yield a $^{206}\text{Pb}/^{238}\text{U}$ weighted mean crystallisation age of 698.1 ± 7.5 Ma (MSWD = 0.21; Fig. 7b). The same analyses yield a concordant age of 701.4 ± 6.2 Ma (MSWD = 2.3; Supplementary Data, Table S3). Older zircon grains are interpreted as antecrysts likely recording prolonged magmatic activity between ca. 745–700 Ma.

Ten Hf analyses on zircon with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 738–677 Ma yield $\epsilon\text{Hf}_{(t)}$ values from $+8.4 \pm 1.0$ to $+10.5$

± 1.2 (mean value of $+ 9.7 \pm 1.0$; Fig. 8a; Supplementary Data, Table S4). The lowest $\epsilon Hf(t)$ values occur in zircon with the youngest $^{206}\text{Pb}/^{238}\text{U}$ single dates (e.g., $+8.4 \pm 1.0$ at 677 Ma and $+9.2 \pm 1.0$ at 697 Ma). T_{DM2} ages are 1.05–1.00 Ga (Fig. 8b).

4.3.2.2. TR21 quartz-monzodiorite. Recovered zircon grains form a single population similar to TR28, with less frequent alteration and recrystallisation.

Forty-one U-Pb isotope analyses were performed on forty-one individual grains (Supplementary Data, Table S3). U concentrations (<170 ppm except for three analyses at 461, 274 and 194 ppm) and Th/U ratios (~ 0.6 – 1.0) are consistent with magmatic zircon. Two discordant analyses were disregarded (U186 and U202). Concordant $^{206}\text{Pb}/^{238}\text{U}$ single zircon dates spread from 738 ± 16 Ma to 686 ± 14 Ma with consistent overlapping arrays. Considering the nineteen concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates equal or lower than 702 ± 18 Ma from the most euhedral prismatic grains yields a $^{206}\text{Pb}/^{238}\text{U}$ weighted mean crystallisation age of 694.2 ± 3.5 Ma (MSWD = 0.35; Fig. 7c). The same analyses yield a concordant age of 695.1 ± 3.3 Ma (MSWD = 2.7; Supplementary Data, Table S3). Older domains, mostly cores, are interpreted as antecrysts crystallised since ca. 740 Ma. Overgrowths may have formed during the same or subsequent magmatic pulses.

Ten Hf isotope analyses of individual zircon grains with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 735–686 Ma give $\epsilon Hf(t)$ values ranging from $+8.2 \pm 1.1$ to $+9.9 \pm 1.1$ (mean value of $+9.0 \pm 1.0$; Fig. 8a; Supplementary Data, Table S4). As for TR28, the lowest $\epsilon Hf(t)$ values occur in zircon with the youngest concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates (e.g., $+8.2 \pm 1.1$ at 686 Ma). T_{DM2} ages are 1.09–1.01 Ga (Fig. 8b).

4.3.2.3. TR25 diorite. Recovered zircon grains form a single population with characteristics similar to TR21.

Thirty-seven U-Pb data spots were made on as many zircon grains (Supplementary Data, Table S3). One discordant measurement (U97) is disregarded, and another analysis (U96) was excluded because of an anomalously low Th/U ratio of 0.06. Otherwise, U concentrations remain rather low (up to 184 ppm) and Th/U ratios are representative of magmatic zircon (~ 0.3 – 0.9). Zircon $^{206}\text{Pb}/^{238}\text{U}$ single dates steadily record magmatic activity from 747 ± 16 Ma to 682 ± 14 Ma. The five autocrysts with the youngest concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates define a $^{206}\text{Pb}/^{238}\text{U}$ weighted mean crystallisation age of 692.9 ± 6.9 Ma (MSWD = 0.78; Fig. 7d). The same analyses yield a concordant age of 694.2 ± 6.1 Ma (MSWD = 1.0; Supplementary Data, Table S3). Older grains are interpreted as antecrysts formed between ca. 747–695 Ma. The oldest single-zircon dates are from inner grain domains, leaving open whether entire grains are preserved antecrysts lacking younger overgrowths, given only one data spot per grain.

Ten Hf isotope analyses of individual zircon grains with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 747–696 Ma yield $\epsilon Hf(t)$ values ranging from $+9.5 \pm 1.1$ to $+11.7 \pm 1.0$ (mean value of $+10.3 \pm 1.0$; Fig. 8a; Supplementary Data, Table S4). The lowest $\epsilon Hf(t)$ values occur in zircon grains with the youngest concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates (e.g., $+9.5 \pm 1.1$ at 702 Ma and $+9.6 \pm 1.0$ at 696 Ma). T_{DM2} ages span 1.02–0.94 Ga (Fig. 8b).

4.3.2.4. WG03-OC-05 granodiorite. Zircon forms two populations based on morphology and internal textures. All grains are subhedral to euhedral, up to 500 μm long, with length-to-width ratios of 1:1.3 to 1:10. The two populations are separated by the presence or absence of distinguishable cores. When present, cores are slightly rounded, from <20 to 200 μm across, and are brighter than oscillatory-zoned overgrowths in cathodoluminescence images. Such a pattern indicates zircon growth around older chemically distinct antecrystic or xenocrystic cores. Most cores appear to be fragments of earlier zircon grains (anhedral shapes not constrained by inner oscillatory zoning), likely from an igneous origin. Some cores show intense alteration (voids and numerous

inclusions), implying a succession of magmatic and hydrothermal events disconnected from the magmatic event from which neoformed, core-less zircon and outer overgrowths crystallise. The second population displays oscillatory igneous growth zoning.

Of twenty-five U-Pb measurements on twenty-one grains, nine discordant were disregarded (Supplementary Data, Table S3). The first population (rounded cores) has low U concentrations (63, 89, 194 and 439 ppm) and magmatic U/Th ratios (~ 0.3 – 0.6). Their concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates range from 841 ± 17 Ma to 812 ± 19 Ma (Fig. 7e), comparable to WG03-OC-03b diorite (Subgroup 1.1). Neoformed grains and overgrowths around inherited cores have very high U concentrations (mostly above 900 ppm and up to 1464 ppm) and lower magmatic Th/U ratios around ~ 0.2 . Nine rim or neoformed-grain analyses yield concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates spanning 762 ± 18 Ma to 655 ± 15 Ma, with minor overlap (Fig. 7e). Outer growth bands are younger than inner ones. Thus, the ca. 840–810 Ma cores are interpreted as inherited xenocrysts overgrown by ca. 760–650 Ma zircon. The youngest concordant $^{206}\text{Pb}/^{238}\text{U}$ single date of 655 ± 15 Ma is interpreted as the crystallisation age of the WG03-OC-05 granodiorite; older dates and inner bands are antecrysts formed between ca. 760–650 Ma. Three grains with similar aspect to the second population yield $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 570–500 Ma. These are not considered emplacement ages because they post-date deformation in the studied area, whereas the sampled granodiorite is foliated (Fig. 3d; Perret et al., 2021). We therefore interpret these dates as representative of the resetting of the isotopic system and Concordia-parallel Pb loss related to later tectono-metamorphic events.

Four Hf analyses of inherited xenocrystic cores with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 834–812 Ma yield $\epsilon Hf(t)$ values ranging from $+7.4 \pm 1.0$ to $+9.5 \pm 1.0$ (mean value of $+8.5 \pm 0.9$; Fig. 8a; Supplementary Data, Table S4). T_{DM2} ages are 1.22–1.05 Ga (Fig. 8b). Despite slightly lower $\epsilon Hf(t)$ values, this signature is comparable to WG03-OC-03b diorite. Eight Hf analyses of antecrysts or autocrysts with concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates between ca. 762–655 Ma yield $\epsilon Hf(t)$ values ranging from $+6.2 \pm 1.7$ to $+9.6 \pm 0.8$ (mean value of $+8.2 \pm 1.0$). As for the three previous samples, the lowest $\epsilon Hf(t)$ values occur in zircon with the youngest concordant $^{206}\text{Pb}/^{238}\text{U}$ single dates (e.g., $+6.2 \pm 1.7$ at 655 Ma and $+6.9 \pm 0.9$ at 684 Ma). T_{DM2} ages are 1.14–1.03 Ga.

4.3.3. Group 2

4.3.3.1. WG03-OC-01b monzogranite. Recovered zircon grains form two distinct populations. A very rare set of grains is rounded to anhedral, fractured, relatively bright in cathodoluminescence, and lacks clear internal zoning, possibly due to alteration; similar features occur as cores in complex zoned grains, suggesting a xenocrystic (or antecrystic) origin. The second population comprises subhedral to euhedral neoformed grains or zircon overgrowths around cores, up to 200 μm long, with length-to-width ratios of 1:2 to 1:4, showing oscillatory igneous growth zoning. This indicates growth as neoformed grains or overgrowths during one or several magmatic pulses; they are antecrysts or autocrysts.

Out of the thirty-one U-Pb isotope measurements made on twenty-nine grains, nineteen discordant were disregarded (Supplementary Data, Table S3). The first population shows <125 ppm U concentrations and Th/U ratios >0.8 compatible with magmatic origin and later alteration, as observed texturally. Two concordant datapoints represent this population, with $^{206}\text{Pb}/^{238}\text{U}$ single dates at ca. 826 Ma and ca. 745 Ma, within the crystallisation age ranges of Subgroups 1.1 and 1.2, respectively. These occur as sparse isolated grains and as cores overgrown by discordant rims and are interpreted as inherited xenocrysts. Neoformed grains and overgrowths have 107–1103 ppm U concentrations and Th/U ratios ~ 0.4 – 0.5 , confirming their magmatic origin. They are interpreted as autocrysts. Ten concordant datapoints yield

$^{206}\text{Pb}/^{238}\text{U}$ single zircon dates between ca. 628–606 Ma and a $^{206}\text{Pb}/^{238}\text{U}$ weighted mean crystallisation age of 620.3 ± 4.5 Ma (MSWD = 0.87; Fig. 7f). The same analyses yield a concordant age of 620.7 ± 3.5 Ma (MSWD = 0.13; Supplementary Data, Table S3). Four younger concordant grains with fractured and recrystallised aspects likely record ca. 560–500 Ma tectono-metamorphic events, as described for WG03-OC-05.

No Hf analyses were performed on inherited xenocrysts. Five Hf analyses on ca. 625–617 Ma autocrusts yield the lowest, yet still highly positive, $\varepsilon\text{Hf}(t)$ values reported in this study, from $+8.2 \pm 0.9$ to $+8.8 \pm 0.8$ (mean value of $+8.5 \pm 0.8$; Fig. 8a; Supplementary Data, Table S4), with T_{DM2} ages of 1.02–0.99 Ga (Fig. 8b).

4.4. Sr and Nd isotope whole-rock analyses

4.4.1. Subgroup 1.1

The initial Nd isotopic composition $\varepsilon\text{Nd}(t)$ for Subgroup 1.1 granitoids emplaced between ca. 840–810 Ma ranges from $+3.8 \pm 2.1$ to $+4.7 \pm 2.0$ (Fig. 8c). Nd- T_{DM2} two-stage model ages span 1.12–1.06 Ga (Fig. 8d), which is consistent with values obtained from Hf isotopes of zircon autocrusts and antecrusts. The single valid initial Sr isotopic ratio ($^{87}\text{Sr}/^{86}\text{Sr}$)_i obtained is 0.702723 ± 0.000061 (Supplementary Data, Table S7).

4.4.2. Subgroup 1.2

Subgroup 1.2 diorite, quartz-monzodiorite and granodiorite emplaced between ca. 760–655 Ma have $\varepsilon\text{Nd}(t)$ values that range between $+3.7 \pm 0.7$ and $+6.2 \pm 1.2$ (Fig. 8c), while ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ranges from 0.702618 ± 0.000802 to 0.702967 ± 0.000215 (Supplementary Data, Table S7). $\varepsilon\text{Nd}(t)$ values tend to decrease for younger intrusions. Nd- T_{DM2} ages span from 0.96 to 0.84 Ga (Fig. 8d), which is younger than the Hf model ages from zircon (1.09–0.94 Ga; Fig. 8b).

4.4.3. Group 2

Group 2 intrusions emplaced between ca. 625–617 Ma have $\varepsilon\text{Nd}(t)$ values ranging from $+3.3 \pm 1.6$ to $+4.6 \pm 1.0$ (Fig. 8c). Related Nd- T_{DM2} ages are 1.01–0.91 Ga (Fig. 8d), matching model ages calculated from the Hf isotopic composition of zircon in the dated sample (1.09–0.94 Ma). ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ranges from 0.701772 ± 0.002517 to 0.702657 ± 0.000917 (Supplementary Data, Table S7).

5. Discussion

We integrate here the geochemical, geochronological, and isotopic record of pre- to syn-tectonic (Group 1, subdivided into Subgroups 1.1 and 1.2) and post-tectonic (Group 2) plutonic rocks of the Kerat domain to define the successive magmatic stages of the Kerat intra-oceanic arc terrane. Based on these results and isotopic data compiled from the literature, we propose a revised geodynamic interpretation that shifts the position of the arc-continent suture zone against the Saharan metacraton further west relative to its commonly accepted location in the studied area (e.g., Abdelsalam et al., 1998; Bailo et al., 2003; Fritz et al., 2013; Abdelrahman et al., 2017).

5.1. Juvenile intra-oceanic arc magmatism between ca. 840–810 Ma

Subgroup 1.1 plutonic rocks comprise foliated, mafic to intermediate intrusions affected by syn-collisional thrusting and subsequent strike-slip shearing (Figs. 2, 3b; Perret et al., 2021). These rocks therefore predate collision and inform the early magmatic evolution of the Kerat domain.

They are enriched in fluid-mobile elements such as LILE, shown as positive concentration spikes in Rb, K, Pb and Sr in spider diagrams (Fig. 5; Supplementary Data, Table S1). This is interpreted as resulting from metasomatism of the mantle source region by hydrous fluids or melts (Stern, 2010). They also display metaluminous I-type signatures and flat HREE

patterns, the latter being inconsistent with residual garnet in the mantle source region (Figs. 4, 5a, d; Supplementary Data, Table S1). Finally, their isotopic compositions are juvenile and suprachondritic, with $\varepsilon\text{Hf}(t) \sim +9.1$ to $+10.7$, ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ~ 0.7034 – 0.7038 (with one lower value at 0.7027) and $\varepsilon\text{Nd}(t) \sim +3.8$ to $+4.7$, with crustal residence times of ~ 200 – 300 Myr (Fig. 8). These observations indicate derivation from a largely juvenile, mantle-derived Neoproterozoic magma reservoir with limited involvement of older continental crust. Collectively, these features are typical of an intra-oceanic arc setting (Stern, 2010), notably in the ANS (e.g., Eyal et al., 2018; Faisal et al., 2020, 2025).

We therefore interpret ca. 840–810 Ma magmatism in the central Kerat domain to represent the constructional stage of an intra-oceanic arc. This interpretation is independently supported by the dominance of interbedded andesitic flows, volcanosedimentary rock sequences, and immature detrital sediments over continental input (Perret et al., 2021), which is typical of intra-oceanic arc depositional environments (Calvert, 2011). Although oceanic seafloor remnants within the Kerat domain and along the Atmur-Delgo suture zone have long been recognised, earlier studies have interpreted them as fragments of connected ensialic back-arc basins related to the Mozambique Ocean (Abdel-Rahman, 1993; Schandelmeyer et al., 1994b; Stern, 1994; Abdelsalam et al., 1995; Abdelrahman et al., 2017).

The magmatic period recorded by Subgroup 1.1 plutonic rocks spans ca. 840–810 Ma (Fig. 7a, b) and is also registered by inherited zircon xenocrysts in younger plutonic rocks (see below; Fig. 7e, f). This timing is compatible with previous broad constraints on andesitic volcanism at ca. 880–720 Ma in the Kerat domain and farther west (Meinhold, 1979; Dawoud, 1980; Ries et al., 1985; Abdel-Rahman, 1993; Stern et al., 1994; Ibinoof et al., 2016). The studied Subgroup 1.1 plutons are mid-crustal, plagioclase-dominated granitoids, with positive Sr and Eu anomalies indicative of plagioclase accumulation (Fig. 5e; Chappell and Wyborn, 2004; DeBari and Greene, 2011). These features imply that the Kerat arc had already developed significantly by ca. 840–810 Ma and likely initiated earlier, with the intermediate part to deeper root of the arc now exposed. This interpretation is not particularly surprising, given that island-arc magmatism has been reported from the northernmost ANS as early as ca. 1040 Ma (Eyal et al., 2014; Vardi et al., 2026).

5.2. Arc maturation and magmatic accretion between ca. 760–655 Ma

Subgroup 1.2 plutonic rocks include diorite, quartz monzodiorite, and granodiorite plutons, structurally affected by syn-collisional thrusting and late strike-slip shearing (Figs. 2, 3a, b; Perret et al., 2021). Their distinct geochemical and isotopic characteristics, together with younger ages (ca. 760–655 Ma; Fig. 7b–e, g), mark a separate magmatic period predating collision.

Subgroup 1.2 plutons are metaluminous I-type granitoids plotting within the tonalite-trondhjemite-granodiorite field (Fig. 4). While maintaining an island-arc affinity (Stern, 2010), they are significantly more evolved than Subgroup 1.1 granitoids. They display steeper chondrite-normalised LREE patterns, higher enrichment in LILE, a relative depletion in HFSE, and stronger negative Ti and positive Hf-Zr anomalies (Fig. 5a, b, d, e and f). In terms of isotopic signature, they display lower $\varepsilon\text{Hf}(t)$ and $\varepsilon\text{Nd}(t)$ values (e.g., WG03-OC-05 granodiorite with $\varepsilon\text{Hf}(t) \sim +7.7$ to $+9.0$ and $\varepsilon\text{Nd}(t) = 3.7 \pm 1.0$; Fig. 8a, c).

Comparable Hf model ages (T_{DM2} ages of ca. 0.94–1.10 Ga; Supplementary Data, Table S4) for both arc magmatism periods suggest derivation from a reservoir with a similar juvenile isotopic signature. Several lines of evidence suggest the contribution of reworked arc-crustal components in melt generation for Subgroup 1.2, such as (i) the presence of xenocrystic zircon cores formed at ca. 840–810 Ma (Fig. 7e, f) in these plutons, (ii) $\varepsilon\text{Hf}(t)$ and $\varepsilon\text{Nd}(t)$ values for Subgroup 1.2 plotting in the Hf- and Nd-crustal evolution arrays defined for Subgroup 1.1 plutonic rocks (Fig. 8a, c) and (iii) a larger span for Hf crustal residence times with respect to Subgroup 1.2 (i.e., ~ 200 – 550 Myr; Fig. 8b, d).

Yet, the dataset presented in this study is insufficient to constrain the nature of the crustal material involved in Subgroup 1.2 petrogenesis. Relative to Subgroup 1.1 granitoids, Subgroup 1.2 ones display higher Th/Nb (0.80–2.5 vs 0.06–0.95, outlier excluded) and La/Nb (4.0–7.7 vs 2.8–3.7, outlier excluded) ratios (Supplementary Data, Table S1), parameters commonly used to assess sediment contributions in subduction-related magmatic systems (e.g., Elliott, 2003; Pearce et al., 2005). Taken alone, these ratios may, however, only indicate increased involvement of a crustal component, with no further discrimination between alternative sources, nor quantification of their relative contribution. Resolving the nature of this component would require dedicated source modelling and/or additional source-sensitive tracers, such as zircon O isotopes, which are beyond the scope of the present study.

Subgroup 1.2 granitoids are geochemically and isotopically similar to ca. 740–710 Ma granitoids studied by Bailo et al. (2003) in the central and northern Kerf domain (Figs. 2, 4, 5, 8). Equivalent magmatism also extends farther north into the Atmur-Delgo arc (Supplementary Data, Table S8), and eastwards into the Gabgaba terrane, where two island-arc plutons formed at 773 ± 17 Ma and 752 ± 11 Ma have been documented (Adam et al., 2022). Similar arc magmatism at ca. 690 Ma in the Egyptian Nubian shield (Stern et al., 2020; Gamal El Dien et al., 2021) points to a regional process of progressive magmatic thickening and crustal maturation leading to arc terrane consolidation by ca. 650 Ma. This coincides with the amalgamation of ANS oceanic arc-derived terranes prior to final collision with the Saharan meta-craton (ca. 750–640 Ma; Abdelsalam et al., 1998). However, Abdelsalam et al. (1998) interpreted the Kerf domain as being shaped by an arc-continent collision event throughout this period, whereas our new geochronological and isotopic data support protracted intra-oceanic arc growth and thickening until at least ca. 655 Ma in its central part. These results therefore call for a reassessment of the tectonic significance and evolutionary history of the Kerf domain (see subsection 5.3).

5.3. Absence of magmatic record in the central Kerf domain between ca. 810–760 Ma

There is a notable absence of concordant $^{206}\text{Pb}/^{238}\text{U}$ single zircon dates between ca. 810–760 Ma in our dataset (Fig. 7g). This apparent magmatic hiatus may simply reflect sampling bias, but it may also be geologically meaningful. If real, such magmatic quiescence in the central Kerf domain could result from (i) the succession of flare-up and lull cycles along magmatic arcs, due to mantle-lithosphere interaction, with tectonic and/or melting drivers remaining debated (Paterson and Ducea, 2015), or (ii) a range of processes that can trigger migration or complete shut-off of the magmatic front. These include flat-slab dynamics, changes in upper plate thickness, slab rollback and trench retreat, terrane accretion, and subduction step-out or polarity reversal (Gianni and Luján, 2021).

It is worth noting that arc magmatism is reported during this time slice in the southern Kerf domain and neighbouring Saharan meta-craton, and even farther south along strike in the Nuba mountains (Küster and Liégeois, 2001; Ibinoof et al., 2016; Evuk et al., 2017). Furthermore, the earlier period of intra-oceanic arc magmatism in the central Kerf domain is mainly expressed in the western part of the study area, whereas younger magmatism is better represented in the east (Fig. 2), in association with a likely thicker arc crust as discussed above. This first-order spatial distribution may therefore indicate protracted eastward migration of the magmatic front through time in the central Kerf area, rather than a purely cyclic, temporal quiescence in arc magmatic activity. If true, this also provides critical insights into the polarity of the subduction zone associated with the Kerf arc, which was likely west-southwest-dipping, considering present-day geometry. We concede, however, that the exact processes involved and the possible diachroneity of arc evolution along strike remain unclear based on available data.

5.4. Post-collisional magmatism and crustal reworking at ca. 620 Ma

Group 2 granitoids are undeformed granodiorite-monzogranite plutons that crosscut all inherited structures and contain xenoliths of earlier intrusive rocks (Fig. 2, 3e, f). They therefore post-date the main collisional and post-collisional deformation that shaped the regional structural framework (Fig. 2; Perret et al., 2021). The dated Group 2 intrusion crystallised at ca. 621.4 ± 4.3 Ma (Fig. 7f), providing a minimum age constraint on the cessation of deformation in the studied area.

The best-studied pluton (WG03-OC-01a, b) exhibits magma-mingling textures (Fig. 3f) between an I-type granodioritic component compositionally similar to Subgroup 1.2 and a monzogranitic component characterised by moderately fractionated REE patterns but comparable LREE contents (Fig. 5c). It also encloses a monzodioritic xenolith resembling Subgroup 1.1 granitoids (Fig. 3f, 4, 5a, d, e), suggesting entrainment of earlier Kerf arc material during magma ascent or emplacement.

Group 2 granitoids are silica-rich (68.6–69.8 wt% SiO_2) and strongly evolved, with very low MgO (0.09–0.77 wt%), low Mg# (5.0–14.7 %), low Cr (7.7–13.8 ppm) and generally low Ni contents (17.4 ppm in WG03-OC-01b, although WG03-OC-01a contains 162 ppm Ni) (Supplementary Data, Table S1). Although such whole-rock compositions may result from extensive fractional crystallisation of primitive mantle-derived magmas, they are also widely regarded as characteristic of evolved felsic magmas and are commonly interpreted to reflect significant crustal involvement during magma genesis or evolution (Collins et al., 1982; Whalen et al., 1987; Frost et al., 2001). Group 2 granitoids also have low initial Sr isotopic ratios ($(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.701772\text{--}0.702657$), positive $\varepsilon\text{Nd}(t)$ values (+3.3 to +4.6), juvenile zircon $\varepsilon\text{Hf}(t)$ values (+8.2 to +8.8) (Fig. 8a, c; Supplementary Data, Tables S4 and S7) and inherited zircon populations (Fig. 7f). Although the exact petrogenetic processes remain uncertain in the absence of more detailed isotopic modelling, the combined geochemical and isotopic evidence is consistent with models involving a crustal component, ranging from limited crustal assimilation to partial or extensive reworking of crustal material.

Several observations suggest that this crustal component was derived from reworking of juvenile Kerf arc crust rather than from involvement of older pre-Neoproterozoic continental crust. First, Group 2 granitoids display trace-element patterns broadly similar to those of Subgroup 1.2 granitoids, differing mainly by stronger enrichment in LILE while retaining comparable HFSE signatures (Fig. 5h). Second, the intrusion contains inherited zircon grains dated at ca. 826 Ma and ca. 745 Ma, similar in age, zoning patterns, and textures to zircon antecrysts and autocrysts from Kerf arc-related plutons (Fig. 7f). Although Hf analyses were not performed on these inherited grains, their ages and textural characteristics suggest incorporation of older Kerf arc material into the ca. 620 Ma magma. Third, Group 2 granitoids yield Hf- and Nd- $T_{\text{DM}2}$ model ages of ca. 1.12–0.99 Ga, comparable to those of Group 1 arc granitoids, while their $\varepsilon\text{Hf}(t)$ and $\varepsilon\text{Nd}(t)$ values plot within the Hf and Nd crustal evolution arrays defined by Group 1 samples (Fig. 8a, c).

These observations do not rule out a juvenile, mantle-derived contribution, but a range of observations indicate that any mantle input was limited or indirect. These include geochemical characteristics or the absence of microgranular mafic enclaves, preventing confident classification of Group 2 plutons as H-type (i.e., hybrid-type; Castro et al., 1991). Additional stable O and H isotope analyses would possibly help to better quantify the distinct source components (e.g., Leake, 1990).

Collectively, these observations favour partial melting and reworking of juvenile Kerf arc crust during post-collisional extension, potentially triggered by lithospheric thinning, delamination, or asthenospheric upwelling, as proposed for other post-collisional granitoid suites across the ANS (e.g., Zoheir et al., 2021; Faisal et al., 2025). The ca. 650–600 Ma interval represents the collision climax and amalgamation of juvenile terranes across the ANS, associated with final

suturing along major arc-arc sutures such as the Nakasib-Bir Umq (e.g., Küster et al., 1993; Schandelmeier et al., 1994a; Johnson et al., 2003; Abdelsalam and Stern, 2017) and Yanbu-Sol Hamed-Allaqi-Heiani (e.g., El-Sayed et al., 2004; Ali et al., 2010; Abdeen and Abdelghaffar, 2011) ocean-ocean sutures. Alternative mechanisms for widespread post-collisional magmatism across the Nubian shield include slab rollback (e.g., Farahat et al., 2011; Flowerdew et al., 2013; Abd El-Rahman et al., 2017) and lithospheric mantle delamination (Avigad and Gvirtzman, 2009; Eliwa et al., 2014; Khalil et al., 2015; Eyal et al., 2018; Megerssa et al., 2020; Hamimi et al., 2022). However, the lack of magmatic diachroneity across the studied area, together with the absence of strong post-collisional extension in the central Keraf domain (Perret et al., 2021), argues against the delamination hypothesis (Wang and Currie, 2015).

5.5. Reappraising the Keraf domain as a structural corridor

Based on newly acquired data, we propose that the central Keraf domain is best assigned to an intra-oceanic arc terrane comparable to those identified further east within the Nubian shield rather than an arc-continent suture zone. A direct implication is that the Keraf shear zone is best interpreted as a feature internal to this intra-oceanic arc terrane, although it did not necessarily develop as a syn-subduction, trench-parallel shear zone and may instead reflect a somewhat younger shear zone that dissected the arc terrane while accommodating final collision between West Gondwana and the ANS.

Indeed the studied area lacks key evidence for a suture zone from data available, including (i) a clearly defined ophiolitic zone preserving remnants, or the totality, of the oceanic lithosphere formed in an intra-oceanic arc or mid-oceanic ridge setting and/or chaotic accretionary mélanges (Haldar, 2020), (ii) rock units recording subduction-related low-temperature high-pressure metamorphism, with peak conditions ranging from blueschist- to eclogite-facies (Brown et al., 2020), (iii) geophysical, stratigraphic, geochemical and isotopic data suggesting the juxtaposition of two distinct lithotectonic blocks (Chetty, 2017). Consequently, although this possibility remains to be rigorously tested by future scale-integrated, multi-disciplinary research, an alternative hypothesis is that the lithospheric domain boundary between the ANS and the Saharan meta-craton may be located further west than traditionally proposed, as suggested by several previous studies (e.g., Abdelsalam and Stern, 1996; Ahmed Suliman, 2000; Küster and Liégeois, 2001; Evuk et al., 2014).

Regarding syn- to post-Keraf intra-oceanic arc deformation, geochronological data collectively suggest (i) a maximum age for transpressive deformation along the Keraf structural corridor of ca. 700 Ma (U-Pb zircon crystallisation age on a deformed granitoid; Evuk et al., 2014), (ii) a minimum age for deformation of ca. 655–630 Ma (U-Pb zircon crystallisation age of discordant intrusions; Evuk et al., 2014 and this study) to 590 Ma (Ar/Ar cooling ages on biotite and hornblende; Küster and Liégeois, 2001). The Keraf structural corridor therefore represents a late sinistral transpressive system accommodating final convergence between the ANS and Saharan meta-craton perhaps as early as ca. 700 Ma, but more likely between ca. 630–590 Ma, coevally with post-collisional denudation and erosion. It is possible, however, that the (re)-activation of segments of the Keraf structural corridor was diachronous, and this interpretation applies first and foremost to the central Keraf domain investigated in this study.

5.6. Implication for the west Gondwana-East African-Antarctic Orogen boundary

The following discussion is based on a compilation of published geochronological and isotopic data from northern Sudan (Supplementary Data, Table S8). Figs. 9–11 show U-Pb and Pb-Pb zircon

crystallisation ages, Sr and Nd isotopes, respectively. While the spatial distribution of available data is uneven, particularly in some poorly exposed regions, the dataset is considered sufficient to identify first-order regional tectono-magmatic trends. Accordingly, the age distribution and isotopic interpolation maps presented in Figs. 9–11 are intended to illustrate broad spatial patterns and potential crustal domains rather than precise terrane boundaries. The resulting interpretations should therefore be viewed as working hypotheses requiring further testing.

This compilation reveals coeval magmatic activity in the Keraf and Bayuda terranes. The Bayudian tectono-magmatic event (ca. 970–890 Ma; Küster et al., 2008) overprinted the eastern edge of the Saharan meta-craton (Fig. 9). If it is well established that the pre-Neoproterozoic crust of the Saharan meta-craton underwent significant reworking during this event (e.g., Ries et al., 1985; Kröner et al., 1987; Rahman et al., 1990; Küster and Liégeois, 2001), its significance in terms of the assembly of the ANS and docking to West Gondwana remains enigmatic (Küster et al., 2008). About 30 Myr later, juvenile intra-oceanic arc magmatism at ca. 860–810 Ma, matching the earliest Keraf magmatic period reported, is also recorded westwards (e.g., Küster et al., 2008; Evuk et al., 2014; Blades et al., 2021; Fig. 9). It remains uncertain, however, whether these occurrences represent a single intra-oceanic arc or multiple multiple, synchronously developed ones. Later granitoids with a juvenile signature dated at ca. 715–630 Ma southwest of the Keraf domain correspond to the second magmatic period established in this study (e.g., Evuk et al., 2014; Fig. 9).

Between ca. 870–650 Ma, isotopic constraints highlight that reworking and/or recycling of Paleoproterozoic to Mesoproterozoic continental crust are restricted to the Saharan meta-craton, west and south of the isotopically juvenile Keraf domain (e.g., Ries et al., 1985; Kröner et al., 1987; Rahman et al., 1990; Küster and Liégeois, 2001; Figs. 10, 11; Supplementary Data, Table S8). This supports the absence of pre-Neoproterozoic basement in the central Keraf domain and its intra-oceanic arc nature. It also maps out the existence of pre-Neoproterozoic continental lithosphere from the Bayuda domain westwards.

Future efforts to unravel Gondwana assembly in its northern sector must therefore integrate the Keraf tectono-magmatic record with the Bayudian event (ca. 970–890 Ma). This includes testing models that could account for the amalgamation of the Keraf intra-oceanic arc onto the reworked Saharan meta-craton margin and to further explore the tectonic nature of the Bayuda domain. At the current level of knowledge, one could speculate that the Saharan meta-craton may have been incrementally reworked by a west-facing island-arc accretionary system, with the ca. 915–900 Ma Bayuda Tectonic Event (Küster et al., 2008) possibly reflecting arc-continent collision, followed by subduction transference (Stern, 2004; Yang, 2022) at ca. 890–860 Ma and development of the Keraf arc system farther to the east. In such a scenario, the ANS-Saharan meta-craton suture zone could lie west of the Bayuda domain, in agreement with global paleogeographic reconstructions (Şengör et al., 2020; Vêrard, 2021; Lom et al., 2024). However, this remains a working hypothesis that requires independent testing through future, multi-disciplinary, geological, geophysical, and isotopic investigations.

6. Conclusions

Geochemical, geochronological, and isotopic data from magmatic rocks across the central Keraf domain constrain its geodynamic evolution between ca. 840–620 Ma and clarify the sequence of tectono-magmatic events and crustal growth involved in the assembly of western Gondwana. The combined evidence in its central part indicates that the Keraf domain is best interpreted as an independent intra-oceanic arc terrane, rather than as the arc-continent suture zone marking the western boundary of the Arabian-Nubian shield. The Keraf intra-oceanic island-arc magmatic evolution comprises two main periods: (i) an early,

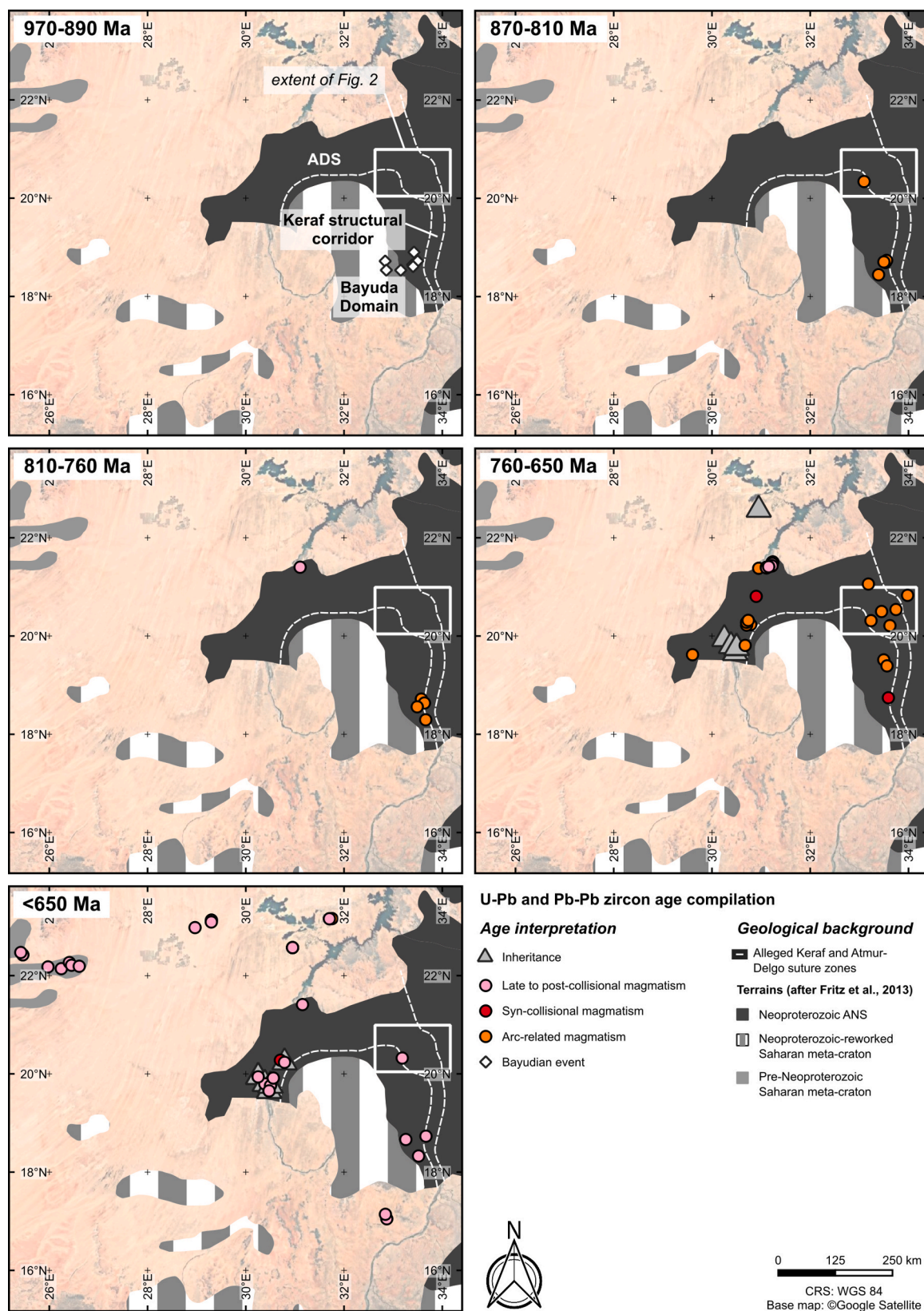


Fig. 9. Space-time evolution of the magmatic record (U-Pb and Pb/Pb zircon crystallisation ages) in the eastern Saharan meta-craton and westernmost Nubian shield. Time slices are those discussed in the text. Data presented are from this study (extent represented by a white rectangle) and literature. The data compilation is provided in [Supplementary Data \(Table S8\)](#). The extent of the map is reported in [Fig. 1](#) and the extent of [Fig. 2](#) is reported on each time slice maps. For reference, the extent of [Fig. 2](#), the Atmur-Delgo suture (ADS), the Kerat structural corridor and the Bayuda domain are labelled for the first time slice map.

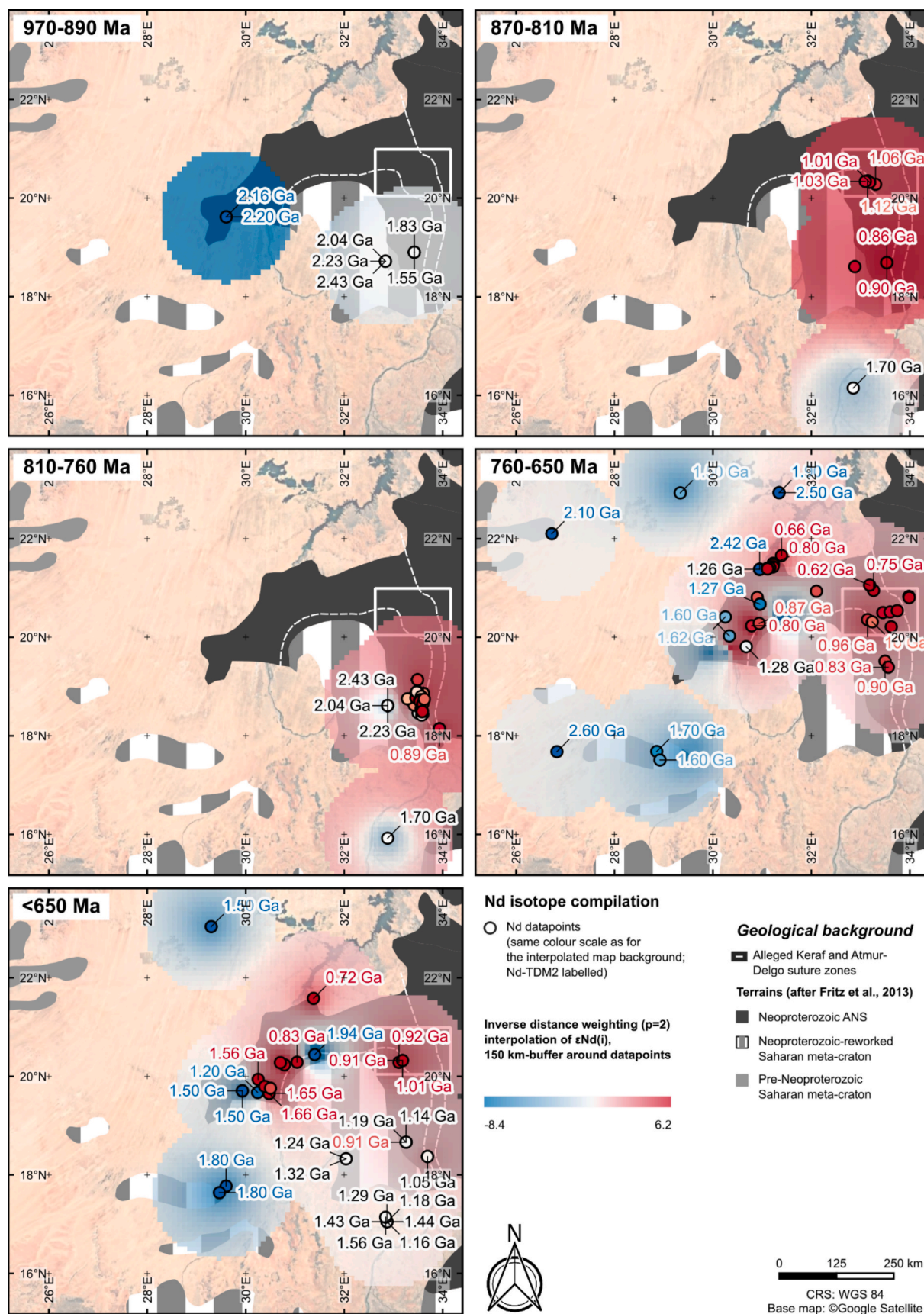


Fig. 10. Space-time evolution of the Nd isotopic record in the eastern Saharan meta-craton and westernmost Nubian shield. See details in Fig. 9 caption.

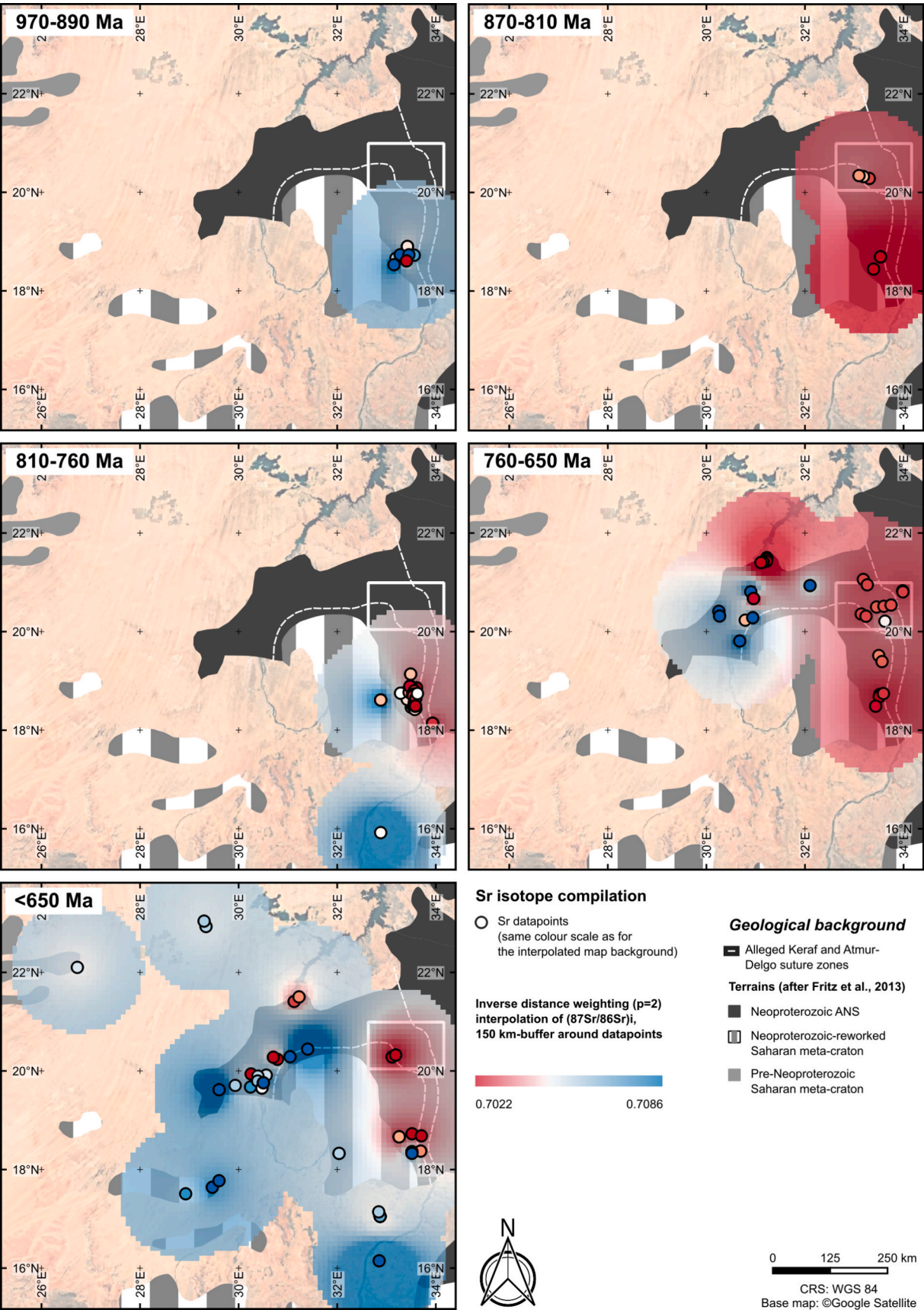


Fig. 11. Space-time evolution of the Sr isotopic record in the eastern Saharan meta-craton and westernmost Nubian shield. See details in Fig. 9 caption.

juvenile intra-oceanic arc phase between ca. 840–810 Ma; and (ii) a protracted stage of mixed-source magmatic accretion between ca. 760–655 Ma, involving mantle-derived magmas and an early arc crustal component. An apparent absence of magmatic record between ca. 810–760 Ma may reflect either sampling bias, tectonic re-organisation, or changes in subduction dynamics and migration of the magmatic front (e.g., slab rollback, flat slab). Finally, post-collisional magmatism at ca. 620 Ma involves assimilation or reworking of the Keraf arc crust, with limited, if any, input from enriched mantle, in a context of regional collision and amalgamation of juvenile terranes across the Arabian-Nubian shield.

A regional isotopic data compilation shows that juvenile magmatism contemporaneous with the Keraf arc evolution is also recorded in the Bayuda domain farther west. It is superimposed on an older Bayudian magmatic event (ca. 970–890 Ma) that strongly reworked pre-Neoproterozoic continental crust along the eastern margin of the Saharan meta-craton. This pattern suggests that the Keraf intra-oceanic arc ultimately accreted onto a reworked continental domain to the west, and that the main Arabian-Nubian shield–Saharan meta-craton suture is more likely to be located west of the Bayuda terrane. Future work should therefore aim to locate and characterise (i) the arc-continent suture zone between the Keraf terrane and the Saharan meta-craton, and (ii) the potential arc-arc suture zone separating the Keraf and Gabgaba-Gebeit terranes, in order to fully integrate the Keraf tectono-magmatic record within the Arabian-Nubian shield assembly and western Gondwana formation.

CRediT authorship contribution statement

Julien Perret: Writing – original draft, Visualization, Validation, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization. **Aurélien Eglinger:** Writing – review & editing, Validation, Supervision, Methodology, Funding acquisition, Formal analysis, Data curation, Conceptualization. **Pierre Bouilhol:** Writing – review & editing, Validation, Investigation, Formal analysis, Data curation. **Quentin Masurel:** Writing – review & editing, Validation, Formal analysis, Data curation. **Anne-Sylvie André-Mayer:** Writing – review & editing, Supervision, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization. **Julien Feneyrol:** Writing – review & editing, Validation, Supervision, Investigation, Funding acquisition. **Andrei Lecomte:** Writing – review & editing, Investigation. **Armin Zeh:** Investigation.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.gr.2026.08.003>.

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