



Review Article

Crustal growth and reworking of the Southern West African craton: Implications for regional tectonics

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ABSTRACT

This contribution reviews the crustal growth of the West African Craton and its geodynamic evolution using whole rock geochemistry together with U-Pb geochronological and Lu-Hf isotopic records preserved in igneous zircon grains from felsic igneous rocks. Our study shows that the crustal growth in the southern West African Craton during the late Siderian and most of the Rhyacian (~2350-2300 Ma and ~2300-2060 Ma, respectively) was largely driven by the opening and subsequent closure of a large oceanic basin that fragmented the Neoproterozoic continental lithosphere. This tectonic evolution involved two major orogenic cycles including the Eoeburnean (~2285-2120 Ma) and Eburnean (~2120-2095 Ma) orogenies, each marked by distinct pulses of both juvenile magmatism and associated crustal reworking. Geochemical, geochronological, and isotopic data acquired on Eoeburnean felsic igneous rocks are used to investigate the process leading to crustal growth and its evolution during the Rhyacian. In central and western Ghana, central Cote d'Ivoire, and Burkina Faso, melt derivation from sources where hornblende and/or garnet were stable is consistent with crustal thickening and the injection of mantle derived magmas. In contrast, eastern and southeastern portions of the sWAC record reworking of Archean crust, as indicated by negative ϵ_{Hf} values, whereas the western Kénéma-Man Domain over in Liberia and Sierra Leone exhibit intact Archean continental crust. During the Eburnean orogeny, more extensive crustal melting occurred at both deep and shallow structural levels, with ϵ_{Hf} values indicating both continued juvenile input and reworking of earlier crust, including ongoing reworking of the northern margin of the Archean Kénéma-Man Domain. Late-orogenic bimodal magmatism (~2095-2040 Ma) in southern Guinea is interpreted as a response to asthenospheric upwelling, likely triggered by lithospheric delamination or slab break-off following arc-continent collision. These magmatic rocks exhibit mixed isotopic signatures, consistent with contributions from both mantle-derived and reworked crustal sources. We propose that the tectonic evolution of the southern West African Craton was controlled by at least two partly synchronous subduction zone segments. The first was active during the Eoeburnean orogenic cycle and was southeast-dipping beneath the Archean BurGhana block in the eastern part of the craton. The second one was active during the Eburnean orogenic cycle and east-dipping, likely extending northwards into the Reguibat Domain and extended in the southern West African Craton where Precambrian terrains accreted onto the Archean Kénéma-Man Domain. We

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propose that the convergence and diachronous collision of intra-oceanic arcs with Archean lithospheric blocks (Kénéma-Man, BurGhana, Reguibat) marked the protracted closure of a ~3,500 km wide oceanic basin.

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1. Introduction

Cratonization of ancient land masses is a fundamental geological process that, in the Archean, contributed to the growth and stabilization of lithospheric blocks that constitute the cores of all continents on Earth (as defined by [Condie, 1998](#)). Crustal growth and its evolution underpins much of the Earth's tectonic history as it involves the thickening, cooling, and chemical differentiation of the Earth's crust and upper mantle, leading to the development of regions that are tectonically stable and resistant to deformation over billions of years ([Condie, 1998](#); [Griffin et al., 2003](#); [Rudnick, 1995](#)). Most models suggest that the generation of continental crust involves at least two stages of differentiation: i) the extraction of basaltic magmas from the mantle followed by ii) fractional crystallization of these mafic magmas or partial melting of their metamorphosed product, i.e. amphibolites (e.g. [Moyen, 2011](#); [Rudnick, 1995](#)). The process of progressive maturation of a mafic lower crust into felsic continental crust varied both in time and from craton to craton, with the overall argument suggesting that the rates of continental crust formation were faster in the early Earth when compared to the Phanerozoic due to progressive secular cooling of the mantle and the

crust ([Hawkesworth et al., 2024](#); [Hawkesworth and Kemp, 2006](#)).

Recognition of the geodynamic engine(s) underlying crustal growth processes is often hindered by protracted and complex tectono-thermal evolution of Precambrian terrains. Nd isotopic and U-Pb and Lu-Hf isotopic systematics acquired on zircon offer a reliable tool with which to peer into the processes that led to the emergence and stabilization of continental masses ([Amelin et al., 2000](#); [Amelin et al., 1999](#); [Belousova et al., 2010](#); [Condie, 2013](#); [Condie and Aster, 2013](#); [Condie et al., 2011](#); [Harrison et al., 2005](#); [Hawkesworth et al., 2010](#); [Parra-Avila et al., 2018](#); [Parra-Avila et al., 2017](#); [Petersson et al., 2016](#); [Wang et al., 2023](#); [Zeh et al., 2014](#)). The U-Pb dating method is widely utilized to determine the crystallization ages of zircon, while Hf isotopic analysis provides insights into the source and evolution of the crustal material, distinguishing between juvenile (newly extracted from the mantle), reworked (previously existing), and mixing trends ([Spencer et al., 2020](#)). Reworking is particularly significant in understanding the evolution of continental crust ([Vervoort and Kemp, 2016](#)).

Many studies suggest that the bulk of the continental crust extraction (~50-70%) took place during the Archean and continued during the dawn of the Proterozoic ([Armstrong, 1991](#); [Belousova et al., 2010](#);

Dhuime et al., 2012; Hawkesworth et al., 2010). This period of the Earth's history is also critical, as it may mark a significant transition from Archean-specific tectonic modes (e.g. stagnant or sluggish lid; Lenardic et al., 2008) to the emergence of modern plate tectonics (e.g. mobile lid; Lenardic et al., 2008). Multidisciplinary data suggests that plate tectonics began to operate effectively by c. 2200 Ma (e.g. Brown et al., 2022; Spencer et al., 2021). The Paleoproterozoic ("Birimian") terrains of West Africa, therefore, provide valuable insights into the tectonic modes operating during the late Siderian and Rhyacian (~2350–2050 Ma), a poorly understood period of Earth's history (Cawood et al., 2022; Hawkesworth et al., 2024). The Paleoproterozoic also witnessed the assembly of the supercontinent Nuna, which is considered the first supercontinent in Earth's history (e.g. Evans, 2013). This assembly was likely preceded by a series of rifting and dispersal events involving earlier supercratons, which set the stage for subsequent complex tectonic interactions (Evans, 2013; Kirscher et al., 2022; Zhou et al., 2024).

In this paper, we present the results of a regional scale effort in isotopic data collection of forty-nine new samples from across the

southern West African Craton (sWAC) complemented by the compilation of extensive data already published on the region. This comprehensive dataset provides a detailed review of the tectonic and geodynamic evolution of the sWAC from the Neoproterozoic to Rhyacian that will ultimately lead us to propose a new tectonic model for the assembly of the sWAC.

2. Geological setting

Over recent years, the sWAC has been the focus of research leading to the production of several regional tectonic and metallogenic syntheses (Eglinger et al., 2022; Grenholm et al., 2019a; Masurel et al., 2022; Perret et al., 2025; Thébaud et al., 2020). As part of the West African Craton (WAC; Fig. 1), this region serves as a key location for the study of the transition from the Neoproterozoic to the Paleoproterozoic eras, primarily due to its protracted geological history and significant crust-forming events. The metamorphic record of the sWAC indicates a shift in the Earth's thermal regime, suggesting the onset of plate tectonics during or after the Neoproterozoic, which is reflected in the geodynamic

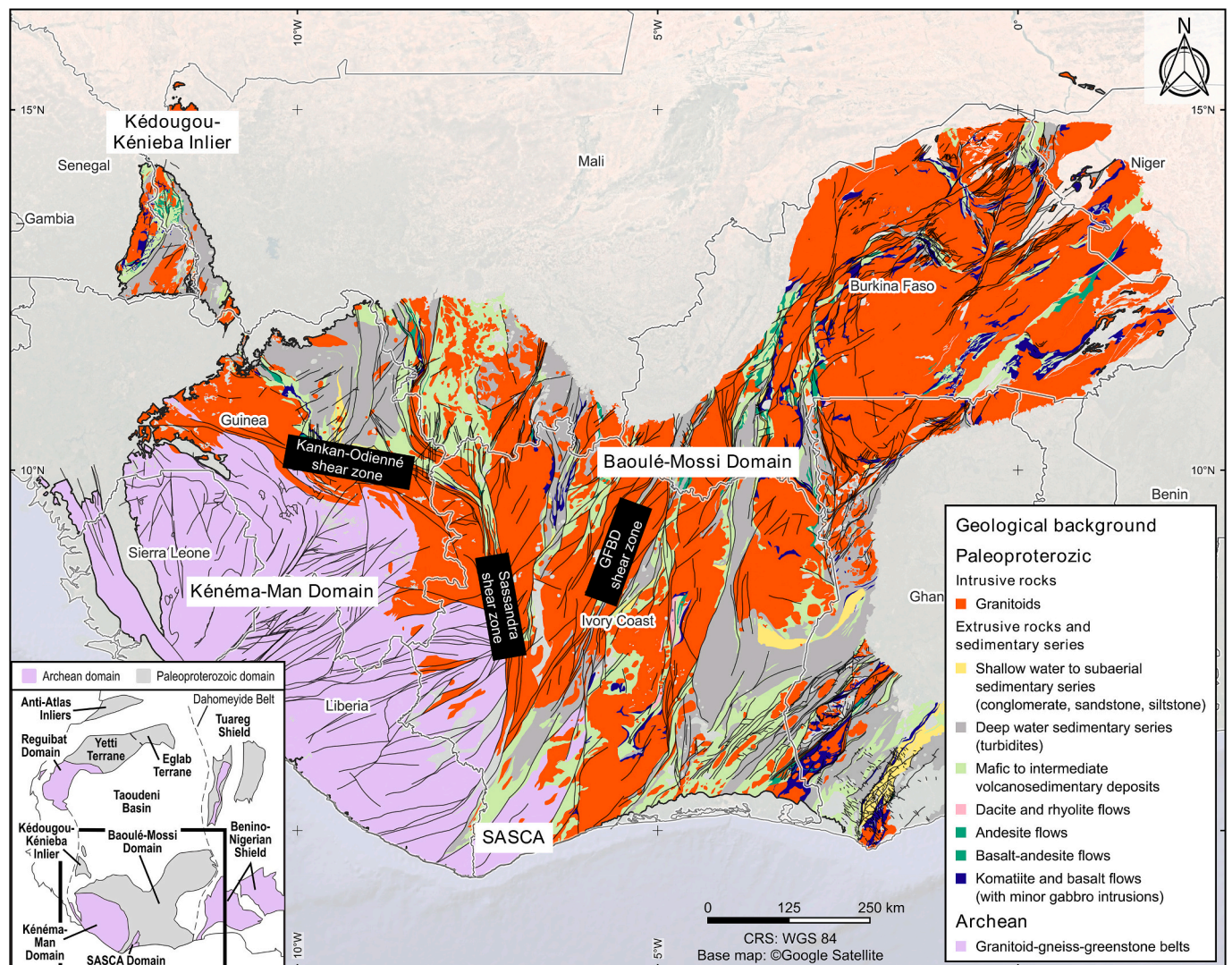


Fig. 1. Geological map of the southern West African Craton (sWAC), including the Paleoproterozoic Baoulé-Mossi Domain and Kédougou-Kénieba Inlier, and Archean Kénéma-Man Domain and Sassandra-Cavally Domain (SASCA) (modified after Perret et al., 2025, and references therein). The Sassandra and Kankar-Odienné shear zones bordering the Archean and Paleoproterozoic domains of the sWAC are labelled for reference. GDFBD shear zone = Greenville-Ferkessedougou-Bobo-Dioulasso shear zone. The bottom-left inset displays the extent of the West African Craton, including the Anti-Atlas Inliers, Reguibat Domain and Yetti-Eglab Terranes to the north of the sWAC. Dashed lines represent the western limit of the WAC, and the Pan-African Dahomeyide belt juxtaposing the WAC and the Archean Benino-Nigerian Shield to the east.

settings interpreted from the Paleoproterozoic rock record (Block et al., 2015).

The WAC geological record illustrates its evolution into a stable craton from the Archean to the Mesoproterozoic. This evolution is punctuated by a prolonged tectonic activity associated with a detailed magmatic, sedimentary and structural record that shaped its current geometry (Ennih and Liegeois, 2008). Notably, the Paleoproterozoic domain within the sWAC is characterized by extensive crust-forming events, including the formation of granitic gneisses and metavolcanic rocks, which are critical for understanding the region's geological evolution (Okonkwo and Ganey, 2012; Sakyi et al., 2020). The presence of Paleoproterozoic mineralization, particularly gold deposits, highlights the economic significance of the WAC and its geological complexity by defining a transitional convergent margin metallogenic cycle (as defined by Huston et al. (2023)) that reflects multiple orogenic periods and a shift in tectonic modes during the Rhyacian in West Africa (Thebaud et al., 2020).

The sWAC encompasses the Archean Kénéma-Man Domain in the west and the Paleoproterozoic Baoulé-Mossi Domain in the east (Auvray et al., 1992; Bessoles, 1977; Milesi et al., 1989; Thieblemont et al., 2004) (Fig. 1). The Kénéma-Man Domain stretches from Cote d'Ivoire and parts of Guinea to Liberia and Sierra Leone, and is associated with tonalite-trondhjemite-granodiorite plutons, gneissic terrains, and greenstone sequences of metavolcanic and metasedimentary rocks that underwent amphibolite to granulite facies metamorphism (Barth et al., 2002; Camil et al., 1983; Gouedji et al., 2014; Koffi et al., 2023; Pitra et al., 2010).

The Archean Kénéma-Man Domain gneisses display U-Pb zircon ages ranging from c. 3600 to 2600 Ma (Gouedji et al., 2014; Milesi et al., 1992; Rollinson, 2016; Rollinson, 2018; Thieblemont et al., 2004), initially interpreted as the result of Leonian (~3400–3000 Ma) and Liberian (~2900–2700 Ma) zircon production periods (Barth et al., 2002; Egal et al., 2002b; Koffi et al., 2020; Milesi et al., 1989; Thieblemont et al., 2004). However, recent research by Rollinson (2016) identified additional zircon production clusters around 3500 Ma, 3260–3050 Ma, and 2960–2850 Ma. These events suggest distinct magmatic pulses, but their details and geodynamic/tectonic implications remain poorly understood. As an alternative, this continuous geochronological record of high-temperature metamorphism and magmatism may also reflect convection of a partially molten Archean crust (Vanderhaeghe et al., 2019).

Inherited zircon cores are often present in younger rocks, especially in boundary regions like the Sassandra-Cavally Domain in Cote d'Ivoire (Koffi et al., 2022) or the Kiniero-Sassandra Domain (Eglinger et al., 2017), where Archean domains are fragmental and amalgamated with Paleoproterozoic terrains (Koffi et al., 2023). These Paleoproterozoic terranes, such as the Ity greenstone belt in Cote d'Ivoire, preserve evidence of felsic igneous activity spanning c. 2150–2060 Ma, recording magmatism both preceding and following the collision between the Kénéma-Man Domain and the Baoulé-Mossi Domain (Traore et al., 2022). The Archean Kénéma-Man Domain was partly destroyed along its northern and eastern margins by tectonic reworking of its crust into the flanking Paleoproterozoic accretionary orogen and crustal reworking of its continental lithospheric mantle into the asthenosphere during hot collision (Chardon et al., 2024; Eglinger et al., 2017; Traore et al., 2022).

The Baoulé-Mossi Domain and Kédougou-Kénieba Inlier span Senegal, Burkina Faso, Mali, Guinea, Cote d'Ivoire, Ghana, Niger, and the northwest corner of Togo and are associated with rocks emplaced during the Rhyacian (Fig. 1; Abouchami et al. (1990); Boher et al. (1992); Egal et al. (2002b)). Its stratigraphic, plutonic, structural, metamorphic, and metallogenic records have been recently reviewed in detail (Perret et al., 2025). Greenstone belts contain volcanic and associated volcano-sedimentary rocks dated between c. 2350 and 2150 Ma (Hayman et al., 2023). Overlying these belts are basins predominantly filled with siliciclastic sedimentary rocks and subordinate volcanoclastic

rocks deposited between c. 2135 and 2110 Ma, along with restricted molasse basins deposited between c. 2110 Ma and 2095 Ma (e.g. Davis et al., 1994; Grenholm et al., 2019b; Hayman et al., 2023; Lebrun et al., 2016; Parra-Avila et al., 2017; Pigois et al., 2003).

These stratigraphic subdivisions delineate the duration of the two main Paleoproterozoic orogenic cycles. The Eoeburnean cycle (~2285–2120 Ma) is proposed to have peaked in region-wide crustal thickening and peak metamorphism at c. 2135 Ma (Grenholm et al., 2019a; Masurel et al., 2022; Perret et al., 2025). The subsequent Eburnean cycle, commencing with widespread basin formation between approximately 2120 Ma and 2095 Ma, and continued with tectonic accretion, basin inversion, and metamorphism (Grenholm et al., 2019a; Masurel et al., 2022). The Eburnean cycle culminated through continued horizontal shortening associated with strike slip shearing interpreted to reflect transpression between c. 2100 Ma and 2095 Ma (e.g. Baratoux et al., 2011; Hein, 2010; Lebrun et al., 2017; Oliver et al., 2020; Perrouty et al., 2012).

The docking of the Archean Kénéma-Man Domain with adjacent Paleoproterozoic domains occurred at c. 2095 Ma, followed by post-collisional high-K volcanism, granitoid plutonism, extension, and exhumation of amphibolite to granulite facies rocks in the eastern sWAC (Eglinger et al., 2017; McFarlane et al., 2019). Throughout the western and southern parts of the sWAC, local areas experienced reverse, normal, or strike-slip shearing/faulting between c. 2095 Ma and 2040 Ma, coeval with high-K magmatism (e.g. Allibone et al., 2020; Masurel et al., 2017). Post-sWAC docking deformation, usually thought to stop after 2000 Ma, may have lasted until c. 1870 Ma, according to recent investigation of the transition zone between the two domains, but no magmatic activity seems to relate to this late deformation (Perret et al., 2026 and references therein).

Prograde metamorphism typically exhibits regional greenschist facies conditions (Baratoux et al., 2024; Hirdes et al., 1992; Kribek et al., 2008; Perret et al., 2025; Taylor et al., 1992) but locally reaches amphibolite or granulite facies (Block et al., 2015; Ganne et al., 2014; John et al., 1999; Pons et al., 1995; Soumaila and Garba, 2006). Voluminous granitoids intruded into the supracrustal rock packages successively between c. 2260 Ma and 2050 Ma (Parra-Avila et al., 2017 and references therein). The granitic rocks that punctuate the deformation history affecting the area display dominantly Paleoproterozoic juvenile Nd and Hf isotopic signatures. Recent studies, however, highlight the local occurrence of reworked ancient crust, suggesting the contribution of other Archean crustal domains (e.g. Petersson et al., 2018).

3. Methodology

Felsic igneous rocks (intrusive and extrusive) were collected along regional traverses to evaluate the crustal architecture and tectonic evolution of the sWAC. Samples were collected for whole-rock geochemical analysis and coupled with in-situ U-Pb and Hf analysis of zircon. A detailed description of the methodology applied to this study can be found in Electronic Supplementary Material 1 (ESM1). Collectively, the assembled database combines newly acquired data together with published dataset spreading from Ghana to the KKI in the north-western corner of the study area.

3.1. Whole-rock geochemistry

Approximately 1.5 kg of least altered, homogeneous rock was collected per sample. Rust stains were cleaned with abrasives, and weathered surfaces were removed. Thin sections were prepared from rock slabs at the University of Western Australia and studied using optical microscopy at the School of Earth and Oceans, Perth. Igneous rock classification followed IUGS systematics (Streckeisen, 1974) and was determined using hand specimen description as a proxy to rock type and modal composition estimate in thin section. Rock samples were crushed, and chips pulverized using low chrome steel vessels. Rock powders were

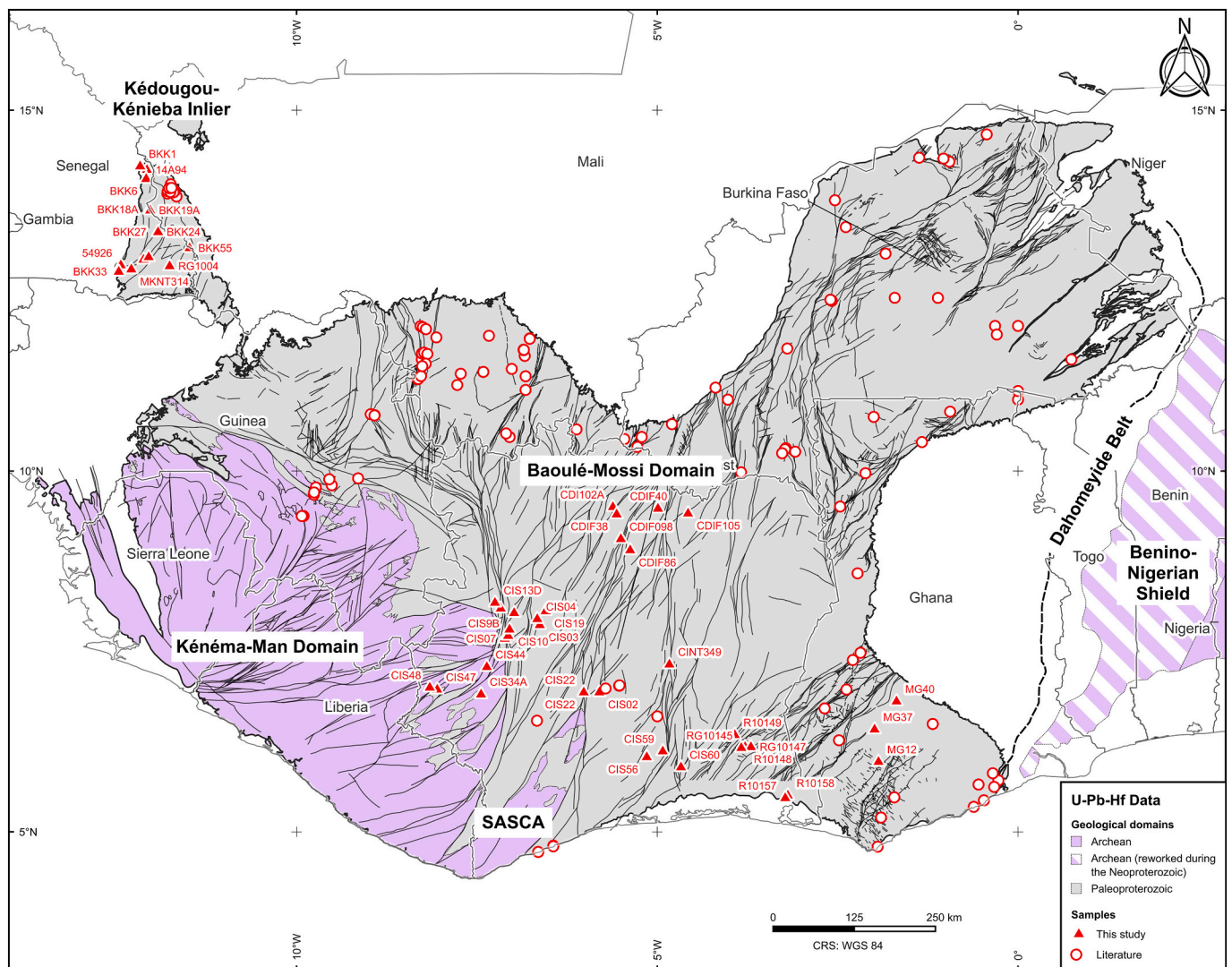


Fig. 2. Distribution of samples collected during the study and legacy U-Pb-Hf data (see ESM2 for data reference). Refer to the legend in Fig. 1 for details on the background geological map.

mixed with lithium metaborate/tetraborate and prepared as a fusion for lithochemistry package LITH-205X analysis at Intertek Laboratories in Perth. Major elements were measured by XRF spectrometry, whereas rare earth elements (REE), high-field strength elements (HFSE) and trace elements (e.g. Ba, Cr, Cs, Ga, Rb, Sc, Sn, Sr, U, V, W) were analyzed by inductively coupled plasma mass spectrometry (ICPMS) and inductively coupled plasma optical emission spectrometry (ICP-OES) at Actlabs in Perth. The Geological Survey of Western Australia Kerba granite standards (Morris, 2007) was analyzed during each session to monitor accuracy and instrument stability. The accuracy of the data was typically better than 3 % for major elements, and trace elements were within 6% of reference values.

3.2. Radiogenic isotopes

Fifty-six samples were analyzed for U-Pb and Hf isotopes during this study, these are displayed on the regional map, which also shows the location of pre-existing and published U-Pb data analyses performed on samples that were also utilized for Lu-Hf characterization (Fig. 2).

3.2.1. Mineral separation and electron microscope imagery

Zircon grains were separated from 1-5 kg rock samples powdered with a jaw crusher and a roller mill and concentrated using heavy liquids

(bromoform: density = 2.84 g/cm³; diiodomethane: density = 3.31 g/cm³) and a Frantz isodynamic magnetic separator. Depending on the sample suites, samples were separated at the University of Lorraine, Nancy (France) or through Geotrack International Pty. Ltd., Melbourne (Australia). Finally, individual grains were handpicked, mounted in epoxy discs, and polished to expose their cores at the University of Lorraine, Nancy (France) or at the University of Western Australia, Perth (Australia). Prior to in-situ isotope analyses, the internal structure of each grain was characterized using cathodoluminescence (CL) and backscattered electron (BSE) imaging electron microscopy with a JEOL JSM7600F instrument with Gatan MiniCL at the University of Lorraine, Nancy (France) or a Tescan Vega-3 at the Centre for Microscopy, Characterisation and Analysis of the University of Western Australia, Perth (Australia).

3.2.2. U-Pb geochronology by secondary ion mass spectrometry (SHRIMP)

Zircon grains for U-Pb analysis were mounted with chips of several U-Pb zircon reference standard materials, i.e., M257 (²⁰⁶Pb/²³⁸U age = 561.3 ± 0.3 Ma, 840 ppm U; Nasdala et al., 2004) or BR266 (²⁰⁶Pb/²³⁸U age = 559.0 ± 0.3 Ma, 903 ppm U; Stern (2001)), as well as crystals of OGC (²⁰⁷Pb/²⁰⁶Pb age 3465.4 ± 0.6 Ma; Stern et al. (2009)) and silicate glass NBS610 (to calibrate common lead peak). U-Pb analyses were performed using the sensitive high-resolution ion microprobe (SHRIMP

II) at the John de Laeter at Centre Curtin University, Perth. Geochronological data was collected using standard operating procedures described by Compston et al. (1984) and Kirkland and Wingate (2012). Operating parameters include a 10 keV primary O₂⁺ beam focused to a ~20 µm diameter spot, a net primary ion current between 1.6 and 3.0 nA and a secondary ion analyzer set to a resolution of ≥ 5000. Six data collection cycles (or scans) were performed per analysis, and count times (per scan) were 10 s for the ²⁰⁴Pb, ²⁰⁶Pb, and ²⁰⁸Pb mass peaks and background and 30 s for the ²⁰⁷Pb mass peak. ²³⁸U, ²⁴⁸Th and ²⁵⁴UO were also collected over six data collection cycles with count times per scan of 5 s, 5 s and 2 s respectively. Analyses of sample zircons were referenced to multiple analyses of the primary and external zircon standards for U/Pb calibration. Calibration and reproducibility uncertainties are included in the errors of ²³⁸U/²⁰⁶Pb ratios and age data. Data reduction was done using the software SQUID v2.5 and ISOPLOT v3.0 (MS Excel add-ins by Ludwig (1999); Ludwig (2003)).

Corrections for common Pb were based on measured ²⁰⁴Pb, assuming an average crustal composition appropriate to the age of the mineral (Stacey and Kramers, 1975). Analytical cut-offs were used to minimize the risk of including zircon analyses with potentially disturbed U-Pb systematics during data interpretation. The reported U-Pb ages were derived from concordant to near-concordant grains (discordance between -3% and +10%). In addition, analyses that comprised more than 1% of non-radiogenic ²⁰⁶Pb and yielded U concentrations over 1000 ppm were rejected. All age estimates are reported at 2σ error and include propagated uncertainties in the standard calibrations. All geochronological plots were generated using Isoplot R (Vermeesch, 2018).

3.2.3. U-Pb and Lu-Hf analyses by LA-ICPMS

In-situ U-Pb and Lu-Hf isotopes analyses carried out by laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS) at Goethe University Frankfurt (GUF), using the same methods and instruments as described in Gerdes and Zeh (2006); Gerdes and Zeh (2009) and Zeh and Gerdes (2012). Laser spots for Lu-Hf (squared spots with an edge length of 40 µm) were placed mostly directly “on top” of the U-Pb laser or SHRIMP spots, both about 20 to 30 µm in diameter, or within the same zone identified by CL imaging.

LA-MC-ICPMS analysis for a subset of Cote d'Ivoire and Ghana samples was performed at the GeoHistory Facility in the John de Laeter Centre, Curtin University, Perth (Australia). Zircon laser targets were selected based on cathodoluminescence (CL) imaging. After two cleaning pulses, grains were ablated using a RESOLUTION incorporating a Compex 102 excimer laser equipped with a Laurin Technik S155 cell. The beam was 22 µm in diameter, the repetition rate was 5 Hz, and the laser energy was 1.8 J/cm² measured at the sample surface. The sample cell was flushed with ultrahigh purity He (320 mL/min) and N₂ (1.2 mL/min). Each analysis consisted of 40 s of background acquisition followed by 20 s of sample data acquisition and another 15 s of background counting. Isotopic intensities were measured using a Nu Plasma II multi-collector ICPMS, using high-purity Ar as the plasma gas. For this work, the following elements were monitored (detector in brackets: H – Faraday Cup; IC – Ion Counter) over an 0.8 s integration time: ²³⁸U (H10), ²³²Th (H8), ²⁰⁸Pb (IC0), ²⁰⁷Pb (IC1), ²⁰⁶Pb (IC2), ²⁰⁴Pb (IC3), and ²⁰²Hg (IC4). Unknowns were bracketed by a suite of standard zircons, with the standard block analyzed every 25 unknowns. The primary reference materials used in this study were GJ-1 (601.7 ± 1.4 Ma; Jackson et al. (2004)), 91500 (1062.4 ± 0.4 Ma; Wiedenbeck et al. (2004)) and OGC, with Plešovice (337.13 ± 0.37 Ma; Sláma et al. (2008)) used as a secondary age standard. ²⁰⁶Pb/²³⁸U ages calculated for all zircon age standards, treated as unknowns, were within analytical uncertainty of the accepted value. The time-resolved mass spectra were reduced using the U-Pb_Geochron4 data reduction scheme in Iolite3.5 (Paton et al., 2010).

Hf isotopic data collected over the Kédougou-Kénieba Inlier and in Cote d'Ivoire and Ghana were collected from selected zircon grains previously analyzed by SHRIMP geochronology. In situ measurements of

Lu-Hf isotopes were carried out using a GeoLas ArF gas (193 nm) laser ablation system coupled to a Thermo Finnigan Neptune Multi-collector ICPMS at the Advanced Analytical Centre (AAC), James Cook University, Townsville (Australia) and a Cetac Analyte G2 laser and Thermo Scientific Neptune PLUS ICPMS at the University of Western Australia. The instrumental setup has been described by Kemp et al. (2009) and Naeraa et al. (2012) for JCU and Petersson et al. (2019).

Data were obtained with a laser fluence of 5–6 J/cm², an ablation rate of 4 Hz, and a laser pulse repetition rate of 4 Hz over a 60 s ablation period. The beam diameter was 30, 40, 50 or 58 µm, depending on the targeted zircon growth phase size. Care was taken to ensure that the analytical site was within the same CL-defined growth domain from which the age data were obtained.

Ablation was carried out in a He atmosphere, with the He carrier gas exiting the cell being combined with Ar in a small volume (3 cm³) glass mixing chamber before transporting into the ICPMS. A small (~0.005 to 0.010 l/min) N₂ flow was introduced into the Ar carrier gas upstream of the mixing chamber to enhance sensitivity (Hawkesworth and Kemp, 2006; Iizuka and Hirata, 2005). The Hf isotope data were derived from a 60 s ablation period, where ablation was ideally conducted as close as practical to any pits resulting from the U-Pb analysis and within the same CL-defined growth phase. Where growth zones of different ages and ¹⁷⁶Hf/¹⁷⁷Hf were intersected during depth profiling, as for the case of drilling through an inherited core into a magmatic rim, the Hf isotope ratios of the two components were isolated from the resulting mixed signal using a customized Microsoft Excel spreadsheet. Only the flattest, most stable portions of the time-resolved signal were selected for integration. Within-run outlier rejection was set at three standard errors of the mean. The correction for the isobaric interference of Lu and Yb on ¹⁷⁶Hf was performed by monitoring the interference-free ¹⁷¹Yb during the analysis and then calculating the magnitude of the ¹⁷⁶Yb interference using the ¹⁷⁶Yb/¹⁷¹Yb ratio of 0.897145 determined by Segal et al. (2003). The much smaller Lu correction was performed similarly by measuring ¹⁷⁵Lu and using ¹⁷⁶Lu/¹⁷⁵Lu = 0.02655 (Vervoort et al., 2004). To correct for instrumental mass fractionation, Yb isotope ratios were normalized to ¹⁷³Yb/¹⁷¹Yb = 1.130172 (Segal et al., 2003) and Hf isotope ratios to ¹⁷⁹Hf/¹⁷⁷Hf = 0.7325 using an exponential law. The mass bias behavior of Lu was assumed to follow that of Yb. Analyses of several reference zircons with different Yb/Hf ratios were regularly interspersed with unknowns during the course of the analytical session and showed sub-epsilon unit reproducibility in ¹⁷⁶Hf/¹⁷⁷Hf (See ESM3). Internal precision on individual analyses presents the within-run analytical uncertainty in interference corrected ¹⁷⁶Hf/¹⁷⁷Hf, calculated on a line-by-line basis. Errors in the initial ¹⁷⁶Hf/¹⁷⁷Hf and εHf incorporate uncertainty in the crystallization age, the ¹⁷⁶Lu decay constant and the measured ¹⁷⁶Lu/¹⁷⁷Hf.

All zircon Hf isotope data (samples and reference zircons Temora-2 and FC-1) are normalized to the solution ¹⁷⁶Hf/¹⁷⁷Hf value of Mud Tank zircon (0.282507 ± 0.000006; (Woodhead and Hergt, 2005), reported relative to JMC 475 ¹⁷⁶Hf/¹⁷⁷Hf = 0.282160) using the laser ablation data generated from this zircon in each analytical session. Analytical uncertainties for individual analyses combine the in-run uncertainty with the reproducibility of Mud Tank zircon analyses from the same session (at 2 standard deviation), added in quadrature.

Chondritic values of Bouvier et al. (2008) were adopted for the calculation of εHf: present day ¹⁷⁶Lu/¹⁷⁷Hf (CHUR, today) = 0.0336 ± 1 and ¹⁷⁶Hf/¹⁷⁷Hf (CHUR, today) = 0.282785 ± 11 (at 2 standard deviation). A ¹⁷⁶Lu decay constant was the one proposed by Söderlund et al. (2004) of 1.867 ± 0.008 × 10⁻¹¹/yr. Two-stage depleted-mantle model ages (TDM2) were calculated assuming the parental magma was derived from average continental crust with ¹⁷⁶Lu/¹⁷⁷Hf = 0.015 (Griffin et al., 2002). The depleted-mantle reservoir was modelled as diverging from a chondritic composition (εHf = 0) at 3.8 Ga (Fisher and Vervoort, 2018), reaching present-day ¹⁷⁶Hf/¹⁷⁷Hf = 0.283238 and ¹⁷⁶Lu/¹⁷⁷Hf = 0.0398 (εHf_DM(0) = +16.0).

Table 1

Geochronological data including both U-Pb and Lu-Hf data over the sWAC from this study and the literature.

Sample	Lat	Long	Group age	Age Error (1 σ)	stdv eHf	eHf average	error	Tdm2 (average)	Source	Data already published
BKK55	13.0878277	-11.483027	2092	11	0.37	5.41	0.72	2.208	This study	
CINT349	7.31761	-4.83262	2179	4	0.34	3.70	0.70	2.376	This study	
14A94	14.171341	-12.073826	2164	8	0.33	4.99	0.74	2.287	This study	
MKNT314	12.79365	-12.29045	2150	8	0.84	5.88	0.75	2.223	This study	
54926	12.85001	-12.43555	2142	5	0.48	5.86	0.76	2.218	This study	
CIS22	6.93087	-6.01887	2135	3	1.13	2.83	2.21	2.396	This study	
RG1009	12.928456	-12.121976	2131	7	1.49	3.10	1.52	2.377	This study	
RG1000	12.96311	-12.049933	2071	5	1.15	3.72	1.49	2.295	This study	
RG1004	12.837805	-11.759491	2072	6	2.74	3.99	1.49	2.279	This study	
CIS59	6.11396	-4.92412	2182	3	1.75	2.23	1.79	2.467	This study	
CIS60	5.8929	-4.67197	2105	7	1.01	3.88	2.72	2.331	This study	
BKK1	14.221988	-12.168141	2080	5	1.12	1.42	1.90	2.440	This study	
BKK6	14.0474653	-12.084251	2093	6	1.12	2.76	1.48	2.369	This study	
BKK33	12.7607498	-12.464764	2145	6	0.79	3.67	1.46	2.353	This study	
BKK18A	13.6217453	-12.039724	2082	7	1.93	6.39	1.79	2.160	This study	
BKK19A	13.6100476	-12.028208	2090	6	0.92	5.94	1.67	2.174	This study	
BKK24	13.3099667	-11.919728	2094	12	0.98	5.82	1.68	2.195	This study	
BKK27	13.3099667	-11.919728	2079	9	0.68	7.04	1.61	2.100	This study	
CIS56	6.03381	-5.1458	2093	6	0.38	3.13	1.02	2.347	This study	
CIS47	6.9698277	-8.0500081	2096	4	0.56	2.97	0.75	2.359	This study	
CIS44	7.28156	-7.36268	3202	13	0.70	1.36	0.79	3.082	This study	
CIS34A	6.90305	-7.44299	2824	7	0.42	0.22	0.62	3.064	This study	
CIS48	6.99764808	-8.1536049	2087	3	0.60	3.20	0.58	2.342	This study	
CIS22	6.93087	-6.01887	2135	3	1.13	2.83	2.21	2.396	This study	
CIS02	6.930915	-5.803378	2097	2	0.81	-1.11	2.24	2.606	This study	
CIS03	7.86426	-6.62898	2095	3	0.78	-1.12	2.25	2.605	This study	
CIS04	7.94696	-6.66418	2110	3	0.68	-1.71	2.23	2.652	This study	
CIS07	7.67439	-7.11326	2752	7	1.32	-9.43	2.16	3.590	This study	
CIS17	8.09242	-7.1698	2803	8	1.02	-6.80	2.05	3.470	This study	
CIS9B	7.71543	-7.06674	2103	2	0.63	-0.42	2.20	2.568	This study	
CIS10	7.8021	-7.04823	2112	2	1.52	0.08	2.20	2.558	This study	
CIS13D	8.17281	-7.25035	2727	6	0.84	-7.75	2.13	3.471	This study	
CIS19	8.05678	-6.54813	2091	2	0.77	-0.67	2.21	2.575	This study	
CIS18	8.0309	-6.98418	2096	3	1.60	-1.92	2.27	2.654	This study	
CDI102A	9.499623	-5.615312	2279	4	0.43	2.43	1.52	2.527	This study	
CDIF098	9.05616	-5.504974	2279	4	1.11	2.62	1.47	2.515	This study	
CDIF40	9.478201	-4.990866	2080	6	1.44	3.18	1.19	2.368	This study	
CDIF105	9.410079	-4.573781	2172	8	0.23	4.53	1.26	2.321	This study	
CDIF38	9.398307	-5.563072	2086	4	0.42	2.05	1.29	2.407	This study	
CDIF86	8.900714	-5.375961	2066	10	0.54	2.06	1.25	2.392	This study	
R10158	5.50165242	-3.1907355	2122	6	0.48	4.48	1.54	2.287	This study	
R10157	5.4661779	-3.2266716	2121	5	0.37	4.54	1.45	2.282	This study	
R10149	6.34004906	-3.9260635	2081	7	0.42	2.30	1.49	2.388	This study	
R10148	6.17347926	-3.7825569	2048	5	0.62	1.76	1.36	2.397	This study	
RG10147	6.17309372	-3.7022983	2055	8	0.33	2.75	1.52	2.348	This study	
RG10145	6.16057252	-3.8354899	2189	14	0.68	3.03	1.39	2.369	This study	
MG40	6.80543	-1.6843	2114	6	0.97	4.80	1.70	2.261	This study	
MG37	6.41701	-1.98864	2091	4	0.76	2.91	1.59	2.358	This study	
MG12	5.96819	-1.93404	2112	5	1.01	6.04	1.66	2.185	This study	
BE13	14.34197	-1.37055	2172	15	0.80	6.24	0.89	2.249	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BE2	14.28757	-0.95172	2126	7	0.76	3.39	0.87	2.390	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BE5	14.33002	-1.03243	2126	11	0.94	5.74	1.01	2.253	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BE6	12.01079	-0.31836	2179	20	1.13	5.88	1.12	2.276	Parra-Avila et al., 2019	U-Pb and Lu/Hf
BF01	13.75246	-2.53802	2120	31	1.98	4.58	1.27	2.316	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF02	13.38028	-2.38712	2145	35	1.03	5.37	0.98	2.282	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF03	13.01286	-1.83645	2130	20	1.68	5.00	1.27	2.293	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF04	12.40040	-1.70917	2114	36	1.21	4.51	1.23	2.311	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF05	12.39939	-1.11051	2138	24	1.47	5.66	1.14	2.259	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF06	12.01079	-0.31836	2179	20	0.40	5.63	0.92	2.164	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF11	14.66517	-0.43407	2193	21	2.20	2.26	0.78	2.505	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF12-01	10.64940	-4.79710	2118	10	2.10	0.33	2.70	2.572	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF7	11.89121	-0.29407	2128	55	0.88	4.06	0.86	2.349	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BF8	11.54592	0.73965	2268	32	0.57	2.12	0.78	2.570	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BNF559	10.44257	-5.45693	2143	5	1.32	1.43	1.05	2.519	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BNF561	10.47148	-5.21599	2120	7	2.26	1.25	3.35	2.519	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BNF562	10.33389	-5.27561	2126	4	3.26	1.75	3.62	2.486	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO13	11.69750	-3.19790	2114	1	1.06	6.34	1.16	2.200	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO253	10.31665	-3.21925	2177	7	1.38	3.15	2.87	2.440	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO257	10.30122	-3.23070	2134	3	0.59	4.38	0.83	2.334	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO261A	10.24650	-3.26962	2195	6	0.56	4.66	0.71	2.362	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO261B	10.24650	-3.26960	2196	6	0.50	4.17	0.64	2.392	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO37A	11.15610	-4.19314	2101	6	0.45	2.42	0.58	2.428	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO425B	10.98940	-4.02143	2107	6	1.17	1.27	0.68	2.502	Parra-Avila et al., 2018	U-Pb and Lu/Hf

(continued on next page)

Table 1 (continued)

Sample	Lat	Long	Group age	Age Error (1 σ)	stdv eHf	eHf average	error	Tdm2 (average)	Source	Data already published
HO425D	10.98940	-4.02143	2100	7	1.84	3.46	0.92	2.364	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO480	12.35803	-2.58101	2165	11	1.10	3.29	0.91	2.423	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO621	9.98252	-3.83905	2121	8	1.06	2.56	0.79	2.434	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO631A	10.26827	-3.08917	2150	9	0.59	5.82	0.91	2.258	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO631B	10.26827	-3.08917	2134	9	1.15	6.12	1.52	2.233	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO640A	12.37364	-2.59993	2169	7	1.10	3.85	1.14	2.392	Parra-Avila et al., 2018	U-Pb and Lu/Hf
HO640C	12.37364	-2.59993	2114	5	1.66	4.17	1.77	2.334	Parra-Avila et al., 2018	U-Pb and Lu/Hf
JM012	5.47760	-1.71465	2191	6	1.54	6.35	1.73	2.256	Parra-Avila et al., 2015 and Parra-Avila et al., 2018	U-Pb and Lu/Hf
JM020	5.19696	-1.89885	2160	8	1.59	6.66	1.76	2.220	Parra-Avila et al., 2015 and Parra-Avila et al., 2018	U-Pb and Lu/Hf
KL274	10.77452	-8.91775	2092	33	2.30	5.66	4.91	2.241	Parra-Avila et al., 2018	U-Pb and Lu/Hf
KL565	10.78729	-8.97258	2094	6	2.48	8.20	3.66	2.077	Parra-Avila et al., 2018	U-Pb and Lu/Hf
KL630	10.77046	-8.91985	2069	14	1.28	1.44	4.79	2.467	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-066	11.19229	-7.77283	2099	6	0.94	1.85	1.59	2.461	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-068	10.47100	-7.04710	2102	7	2.09	3.73	2.78	2.354	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-070	10.52081	-7.09607	2119	8	1.60	1.71	1.43	2.501	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-079B	11.12372	-6.82612	2087	8	1.36	3.09	1.17	2.377	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-080	11.10898	-6.81391	2085	6	2.39	5.61	2.07	2.222	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-086	10.57669	-6.11971	2149	6	1.99	3.25	1.05	2.415	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-105	11.31350	-6.82545	2101	10	1.70	3.46	1.23	2.365	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-107	11.87412	-7.33345	2082	10	2.30	3.13	6.47	2.370	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-113	12.00722	-8.28058	2089	4	1.35	3.97	1.59	2.325	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-114	11.99183	-8.24928	2084	7	2.02	3.94	3.21	2.323	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-116	11.96591	-8.20924	2090	6	1.22	5.39	1.27	2.240	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-117	11.96624	-8.20652	2093	5	1.40	6.50	1.42	2.174	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-150	11.85183	-8.06252	2093	5	1.17	3.99	1.07	2.326	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-177	11.54660	-8.26385	2135	6	0.51	7.50	1.95	2.145	Parra-Avila et al., 2018	U-Pb and Lu/Hf
ML12-178	11.54895	-8.26451	2131	10	2.96	4.60	2.92	2.318	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU001	11.83757	-6.76849	2102	11	0.95	4.92	1.12	2.277	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU002	11.83297	-6.76769	2103	14	1.89	3.79	1.23	2.346	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU003	11.59578	-6.83742	2105	6	3.36	8.59	4.34	2.056	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU004	11.68288	-6.85295	2115	10	1.34	3.64	0.86	2.364	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU006	11.67488	-6.85231	2103	7	0.57	1.47	1.04	2.487	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU007	11.67930	-6.84954	2101	11	0.56	1.20	0.69	2.517	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU012	11.41515	-7.01747	2094	4	2.48	1.89	1.07	2.455	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU015	11.37352	-7.41225	2099	5	1.43	1.90	1.07	2.458	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU016	11.37354	-7.41224	2095	6	2.91	2.52	2.95	2.425	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU021	11.27172	-8.32080	2085	7	2.52	5.07	2.64	2.259	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU022	11.27521	-8.31371	2153	24	2.50	2.31	2.73	2.473	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU023	11.31564	-8.27517	2096	11	1.35	2.06	0.79	2.469	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU029	11.31552	-8.27492	2114	11	1.97	1.68	1.27	2.503	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU034	11.37380	-7.41178	2098	7	0.57	2.09	1.66	2.445	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU037	11.34823	-7.72443	2088	8	3.73	3.80	2.89	2.356	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU051	11.62985	-8.26297	2077	5	0.48	5.28	0.79	2.237	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU052	11.62905	-8.26175	2076	6	1.77	4.82	1.37	2.264	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU054	11.64055	-8.23178	2092	4	1.78	7.07	1.49	2.139	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU056	11.62224	-8.18713	2091	6	1.08	6.53	0.98	2.174	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU057	11.48783	-8.21078	2088	5	0.99	3.93	1.25	2.332	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SU059	11.45288	-8.25838	2128	10	1.18	6.87	0.89	2.192	Parra-Avila et al., 2018	U-Pb and Lu/Hf
NG1	10.75000	-2.00000	2125	20	1.33	2.79	0.97	2.423	Parra-Avila et al., 2019	U-Pb and Lu/Hf
SG5	6.49462	-1.18206	2058	22	0.84	3.57	0.89	2.326	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SG6	5.65514	-0.54738	2084	10	2.01	-1.98	0.95	2.680	Parra-Avila et al., 2018	U-Pb and Lu/Hf
BN132	10.40129	-1.33195	2211	3	0.69	4.68	1.19	2.373	Block et al., 2016a	U-Pb and Lu/Hf
BN90	10.82261	-0.94131	2135	3	0.69	3.48	1.06	2.389	Block et al., 2016a	U-Pb and Lu/Hf
BN119	9.96920	-2.11479	2133	3	0.62	4.16	1.07	2.348	Block et al., 2016a	U-Pb and Lu/Hf
BN270	9.50750	-2.46548	2181	3	0.61	3.06	1.06	2.448	Block et al., 2016a	U-Pb and Lu/Hf
BN241	8.58175	-2.22382	2143	6	0.71	4.69	1.06	2.322	Block et al., 2016a	U-Pb and Lu/Hf
NG1	10.75000	-2.00000	2125	20	1.33	2.79	0.97	2.423	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SG5	6.49462	-1.18206	2058	22	0.84	3.57	0.89	2.326	Parra-Avila et al., 2018	U-Pb and Lu/Hf
SG6	5.65514	-0.54738	2084	10	2.01	-1.98	0.95	2.680	Parra-Avila et al., 2018	U-Pb and Lu/Hf
PK101	5.71216	-0.27116	2126	12	0.89	-2.61	1.00	2.750	Petersson et al., 2016	U-Pb and Lu/Hf
PK102	5.81100	-0.34975	2174	6	1.03	2.37	2.29	2.485	Petersson et al., 2016	U-Pb and Lu/Hf
PK103	5.43642	-0.47287	2139	5	2.74	-3.18	1.01	2.793	Petersson et al., 2016	U-Pb and Lu/Hf
PK105	5.62200	-0.33005	2229	4	0.86	4.91	1.07	2.372	Petersson et al., 2016	U-Pb and Lu/Hf
ASGH003A	5.34598	-0.61380	2125	18	1.40	3.36	1.18	2.388	Petersson et al., 2016	U-Pb and Lu/Hf
ASGH007A	4.79348	-1.94556	2173	12	0.98	3.18	1.93	2.435	Petersson et al., 2016	U-Pb and Lu/Hf
ASGH022A	7.48070	-2.18360	2093	2	0.46	4.28	1.11	2.309	Petersson et al., 2016	U-Pb and Lu/Hf
ASGH022C	7.48070	-2.18360	2092	4	0.71	4.22	0.91	2.312	Petersson et al., 2016	U-Pb and Lu/Hf
JD018	9.80208	-9.51212	2087	3	1.47	1.76	1.10	2.457	Eglinger et al., 2017	U-Pb and Lu/Hf
JD019	9.88562	-9.54684	2082	3	1.19	1.00	1.16	2.499	Eglinger et al., 2017	U-Pb and Lu/Hf
JD071	9.89875	-9.14696	2096	3	0.75	2.21	1.23	2.437	Eglinger et al., 2017	U-Pb and Lu/Hf
JD035	9.77226	-9.73294	2093	2	0.97	-11.50	1.30	3.258	Eglinger et al., 2017	U-Pb and Lu/Hf
JD033	9.67190	-9.76110	2082	3	6.75	-15.65	-0.34	3.494	Eglinger et al., 2017	U-Pb and Lu/Hf
JD034	9.70288	-9.75697	2097	5	5.16	-16.27	1.41	3.542	Eglinger et al., 2017	U-Pb and Lu/Hf

(continued on next page)

Table 1 (continued)

Sample	Lat	Long	Group age	Age Error (1 σ)	stdv eHf	eHf average	error	Tdm2 (average)	Source	Data already published
JD704	9.37646	-9.90380	3094	4	0.64	-2.12	1.22	3.437	Eglinger et al., 2017	U-Pb and Lu/Hf
JD705	9.37541	-9.92790	2876	4	0.58	-3.07	-0.35	3.334	Eglinger et al., 2017	U-Pb and Lu/Hf
SAD085	13.80510	-11.66355	2081	3	0.26	4.91	0.86	2.262	Masurel et al., 2017	U-Pb
STA037	13.87897	-11.68916	2162	9	1.03	5.81	1.10	2.250	Masurel et al., 2017	U-Pb
SAD158 β	13.84616	-11.79141	2147	8	0.41	4.93	1.04	2.310	Masurel et al., 2017	U-Pb
SAD077	13.86289	-11.70562	2166	9	0.99	5.14	0.88	2.294	Masurel et al., 2017	U-Pb
SAD158Grd	13.84548	-11.79042	2142	6	0.60	4.52	1.04	2.333	Masurel et al., 2017	U-Pb
SAD182	13.87753	-11.78580	2090	9	0.63	4.58	0.89	2.289	Masurel et al., 2017	U-Pb
SAD185	13.97400	-11.74460	2083	6	1.03	5.09	1.57	2.249	Masurel et al., 2017	U-Pb
SAD033	13.86780	-11.75044	2066	9	0.42	4.49	0.86	2.281	Masurel et al., 2017	U-Pb
SAD075	13.91976	-11.76508	2071	5	0.33	4.99	0.71	2.250	Masurel et al., 2017	U-Pb
SAD065	13.92391	-11.73352	2074	8	0.37	4.80	1.02	2.264	Masurel et al., 2017	U-Pb
SB092	6.97169	-2.37792	2144	10	0.94	3.71	2.11	2.382	McFarlane et al., 2019	U-Pb and Lu/Hf
SB173b	7.38019	-2.29020	2158	6	1.14	3.98	2.11	2.376	McFarlane et al., 2019	U-Pb and Lu/Hf
SB023	6.71047	-2.68040	2135	7	1.38	3.79	1.86	2.370	McFarlane et al., 2019	U-Pb and Lu/Hf
SB248	6.26808	-2.48187	2155	5	0.82	6.05	1.78	2.248	McFarlane et al., 2019	U-Pb and Lu/Hf
CIS61=SP3	4.72083	-6.65041	3238	4	2.11	-3.28	1.57	3.613	Koffi et al., 2022	U-Pb and Lu/Hf
MON4	4.80868	-6.44095	3328	13	0.64	3.74	1.55	3.255	Koffi et al., 2022	U-Pb and Lu/Hf
SP11	4.79892	-6.44230	3245	4	1.07	1.36	1.40	3.339	Koffi et al., 2022	U-Pb and Lu/Hf
YAD036	7.02472	-5.52583	2109	5	0.31	2.81	1.25	2.410	Meriaud et al., 2020	U-Pb and Lu/Hf
YAO_6A	6.98635	-5.71745	2107	4	0.29	2.07	1.33	2.453	Meriaud et al., 2020	U-Pb and Lu/Hf
YAD051	7.02470	-5.52916	2107	4	0.37	2.47	1.43	2.429	Meriaud et al., 2020	U-Pb and Lu/Hf
YAD042	7.02472	-5.52761	2106	4	0.38	2.65	1.24	2.421	Meriaud et al., 2020	U-Pb and Lu/Hf
YAD09	7.02971	-5.52471	2104	5	0.36	3.02	1.19	2.393	Meriaud et al., 2020	U-Pb and Lu/Hf
CIS62	4.72083	-6.65041	2084	5	5.21	-6.07	1.09	2.925	Koffi et al., 2022	U-Pb and Lu/Hf
HDD2613	6.60000	-5.00000	2180	6	0.36	3.90	0.77	2.402	Masurel et al., 2019	U-Pb
IS24	6.53980	-6.66775	2087	12	1.04	1.45	1.56	2.476	Brou et al., 2022	U-Pb

4. Results

Here, we evaluate the whole rock geochemical composition and radiogenic isotope record of felsic igneous rocks distributed over the sWAC. The data presented combines data gathered from the literature and data acquired during this study. This spatially representative, regional dataset allows for a reliable time-integrated view of the crustal evolution of the sWAC. Due to the large amount of data produced and included in this study, U-Pb and Lu-Hf systematics on zircon are summarized in Table 1. Detailed geochronological reports and raw data acquired for each sample dated during this study are provided in Electronic Supplementary Material 1 (ESM1) and 2 (ESM2). For most samples zircon crystals isolated displayed fairly simple growth zoned magmatic texture (ESM1). Where samples displayed more complex textures, analyses included both cores and rims in order to distinguish possible in-situ dissolution and reprecipitation processes that may have affected the igneous rock post-emplacement. Where complex core-rim textures were identified, the data were interpreted so that the age discussed hereafter are attributed to the magmatic emplacement of the igneous stock. A table containing all U-Pb and Lu-Hf data used in this study is also provided in Electronic Supplementary Material 3 (ESM3). Whole-rock geochemical data presented here are included in Electronic Supplementary Material 4 (ESM4). A compilation of the geochronological data included in this paper is also provided in Electronic Supplementary Material 5 (ESM5).

4.1. U-Pb data synthesis

Combining data generated in this study and a compilation of existing literature data compilation over the sWAC by Parra-Avila et al. (2017) indicates that continuous yet diachronous magmatic activity that affected the craton from c. 2300 to 2000 Ma. This diachronic magmatic evolution within the West African Craton is illustrated in Fig. 3. The migration of the magmatic front was approximated to be 35 km/Myr from east to west (Parra-Avila et al., 2017). In this context, the Rhyacian tectonic evolution of the sWAC magmatic activity appears to be concentrated into two main peaks at c. 2160 Ma and c. 2095 Ma (ESM5 and Fig. 4).

4.2. Geochemistry

Whole rock geochemical data (ESM4) comprise 1,709 whole rock analyses reported in the literature and 36 new analyses produced during this study. Analyses are distributed over the whole craton and dominantly in the Baoulé Mossi Domain. Over the 1,709 analyses presented, 77 correspond to dated samples of felsic plutonic and subvolcanic rocks. Data presented in Figs. 5 and 6 below are colored following time periods marking the tectonic evolution of the sWAC including: 2285 to 2200 Ma (pre-orogenic), 2200 to 2135 Ma (early-Eoeburnean tectonic event), 2135 to 2120 Ma (Eoeburnean tectonic event), 2120 to 2095 Ma (Eburnean tectonic event) and 2095 to 2040 Ma (late-orogenic) allowing the evaluation of the geochemical signature of the intrusions with time.

4.2.1. Major elements

The samples considered in this study are all associated with SiO₂ composition >52 wt %. Analyses were filtered to eliminate samples that were too weathered or hydrothermally altered. The filtering process involved eliminating samples from literature with loss on ignition superior to 3% and/or that have wt.%. SiO₂ > 80%. Harker diagrams of major elements against SiO₂ show relatively simple linear trends suggesting that the samples considered are least altered (Fig. 5). Accordingly, their geochemistry is indicative of their original sources and, potentially, the tectonic environments in which they were generated.

In the plutonic classification diagram proposed by Enrique and Esteve (2019), the samples exhibit major geochemical compositions that plot in the quartz diorite to monzogranite and syenogranite fields, with most sample plotting over the granodiorite and granite fields (Fig. 6). Using the modified alkali-lime index (MALI) (Fig. 7a; Frost et al. (2001) most felsic igneous rocks plot in the calc-alkalic to alkalic-calcic field with only limited samples returning calcic or alkalic compositions. When taking age into account, it is apparent that samples evolve to some extent, from a dominantly calcic composition towards more alkali-calcic compositions. This trend is, however, not systematic.

In a Fe-number vs. SiO₂ diagram (Fig. 7b; Frost et al. (2001)), felsic igneous rocks define a coherent trend. Most samples with <70 wt.% SiO₂ plot in the magnesian field, whereas samples with >70 wt.% SiO₂ are predominantly ferroan. On an alumina saturation plot (Fig. 7c; Barton

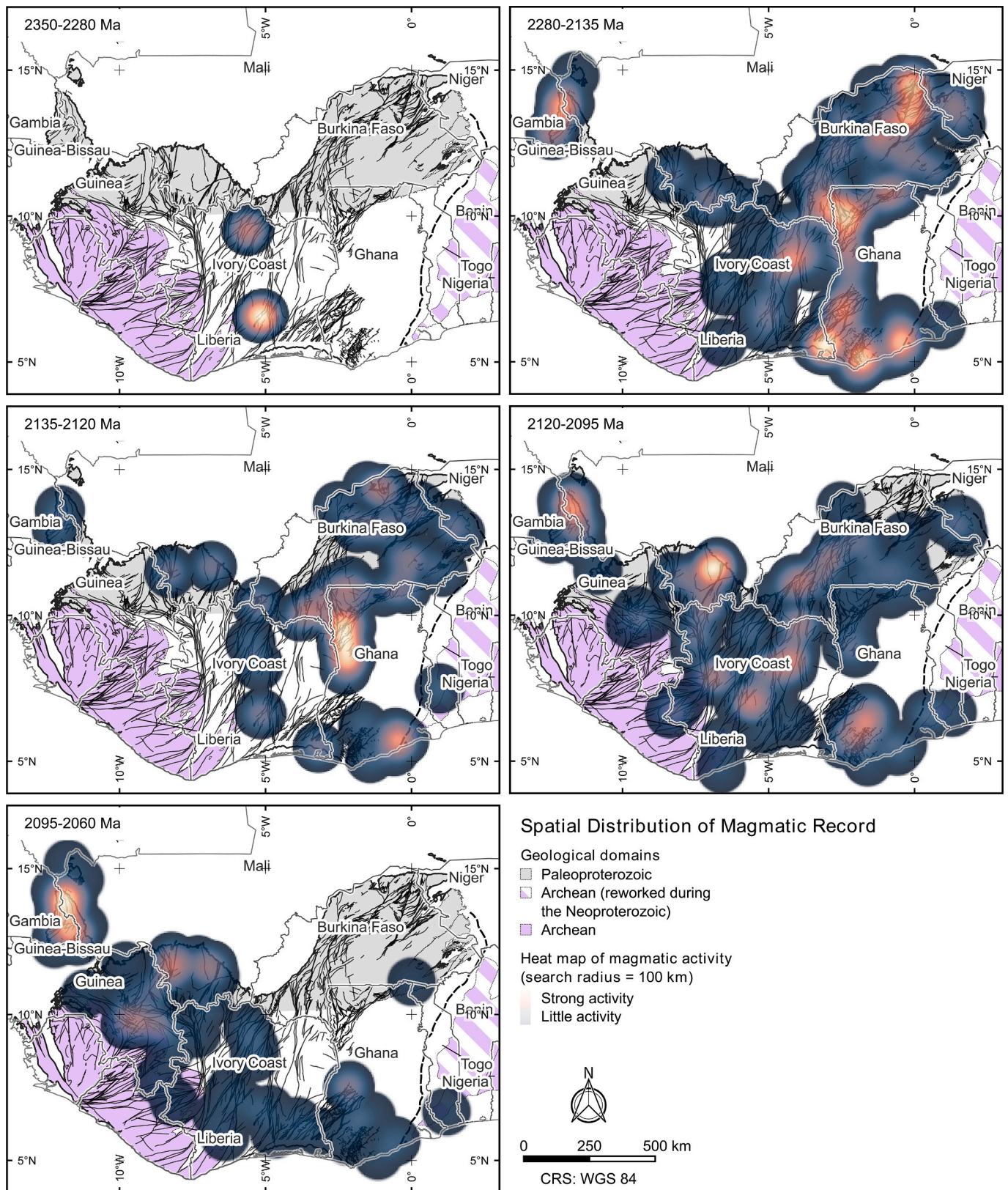


Fig. 3. Heat map of magmatic activity from c. 2300 to 2060 Ma, relative to whole time span. This analysis was undertaken to ascertain whether the apparent spatial variation was due to sampling bias since this study sampling may have reduced Eoeburnean signal. The darker blue domains mean more samples per km² for the period considered. The successive spatial density analyses for five time slices show a clear trend for dominant magmatic activity in the eastern SWAC between c. 2300 and 2110 Ma, followed by a period of evenly spread magmatic activity, and a final period when magmatic activity dominated in the west between c. 2095 and 2060 Ma.

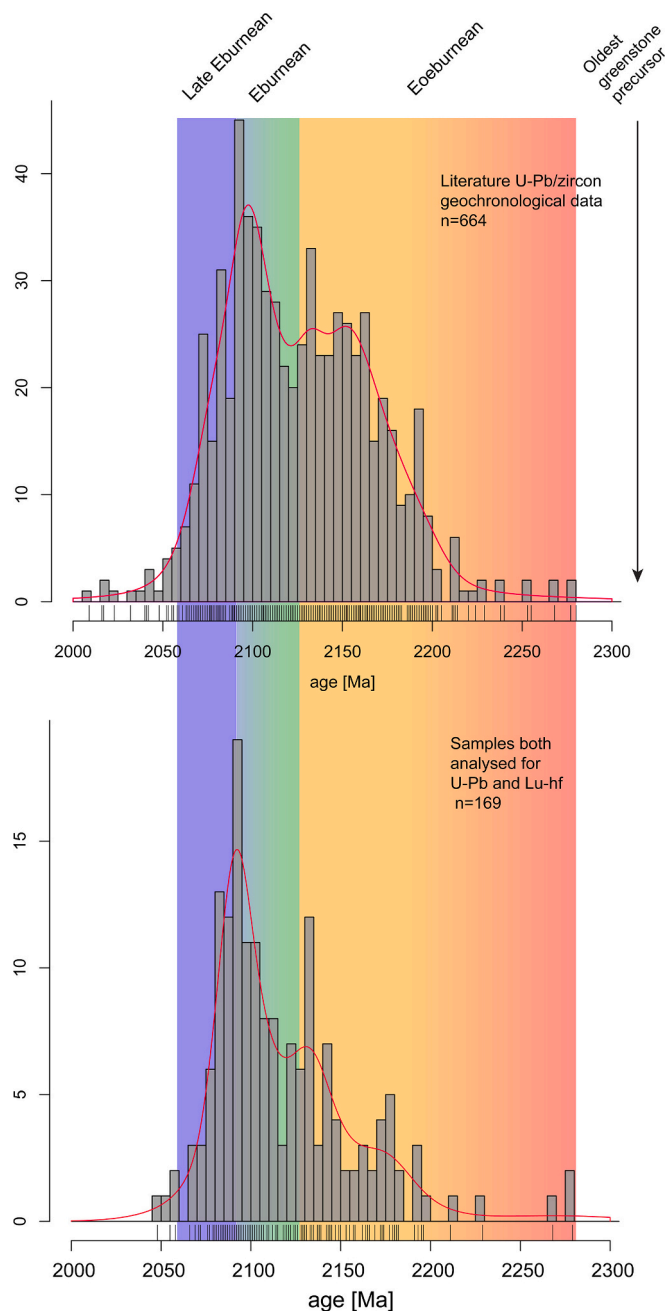


Fig. 4. Probability density plot of U-Pb age data from samples acquired on zircons from this study and the literature (see ESM4 for details).

and Young (2002)), sample with SiO_2 concentration > 63 wt.% show that most samples plot across the metaluminous to weakly peraluminous fields with some samples plotting in the strongly peraluminous field. When considered by age categories, samples emplaced prior to 2135 Ma do exhibit a more metaluminous composition with only a couple of samples plotting in the weakly and strongly peraluminous field. Dated samples from 2135 Ma to 2040 Ma do overlap across the metaluminous, weakly peraluminous and strongly peraluminous field with most samples displaying a weakly peraluminous composition.

In the ternary diagram of Laurent et al. (2014) (Fig. 8), the vast majority of samples plot in the fields of low-K to high-K mafic source rocks, spreading towards the tonalite field. When considered with age, samples emplaced prior to 2120 Ma point towards a dominant mafic source, except for two samples that indicate a tonalitic source. From 2120 Ma onwards, many samples still plot in the mafic source fields

(several indicating a high-K mafic source) or extend through the tonalite field, while a few samples extend into the metasedimentary field. This source evolution is similarly reflected by the $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio. When considering ages, samples analyzed suggest evolution from a dominantly primitive sodic source with a mean value of 3.1, and progressively appear to become more potassic with time, recording ratios of 2.2, 2.0, 1.2 and 1.1 for the early-Eoeburnean, Eoeburnean, Eburnean and late-orogenic stages, respectively (Fig. 9).

4.2.2. Trace elements

Trace element data were explicitly plotted for the subset of 82 analyses for which geochronological constraints exist (Fig. 9). In primitive mantle normalized trace element spider diagrams (Sun and McDonough, 1989), all samples show prominent negative Nb-Ta-Ti anomalies (Fig. 9; bold red) and enrichment in large ions lithophile elements (LILEs) and Th-U (Fig. 9; bold black). Note that these anomalies appear to increase for most of the LILEs, with the potential exception of Ba as time proceeds. Similarly, REE patterns appear associated with a light REE (LREE) enrichment, corresponding to $(\text{La}/\text{Yb})_{\text{cn}}$ ranging from 3.6 to 154 (Fig. 10). The pre-orogenic (2300–2200 Ma), felsic igneous samples present $(\text{La}/\text{Yb})_{\text{cn}}$ ratios between 5.2 and 35.9 with a mean value of 17.17, $(\text{Gd}/\text{Yb})_{\text{cn}}$ ratios between 1.2 and 6.6 with a mean value of 1.9. $(\text{Gd}/\text{Yb})_{\text{cn}}$, Eu/Eu^* ratios between 0.83 and 0.94 with a mean value of 0.89, and Sr/Y values with values from 11.9 to 44.6 with a mean value of 28.8.

The early-Eoeburnean plutons (2200–2135 Ma) present $(\text{La}/\text{Yb})_{\text{cn}}$ ratios comprised between 4.7 and 57.3 with a mean value of 17.2, $(\text{Gd}/\text{Yb})_{\text{cn}}$ comprised between 1.1 and 3.4 with a mean value of 2.2, Eu/Eu^* comprised between 0.56 and 1.6 with a mean value of 0.96, and Sr/Y values with values from 2.48 and 204 with an average value of 65.3.

The Eoeburnean stage (2135–2120 Ma) is associated with igneous felsic rocks showing $(\text{La}/\text{Yb})_{\text{cn}}$ ratios comprised between 35.0 and 77.2 and a mean value of 56.1, $(\text{Gd}/\text{Yb})_{\text{cn}}$ ratio comprised between 3.5 and 4.6 and a mean value of 4.0, Eu/Eu^* comprised between 0.89 and 1.76 with a mean value of 1.2 and Sr/Y values with values from 19.6 and 407 with an average value of 139 (Fig. 10).

The Eburnean stage (2120–2095 Ma) is associated with a range of trace element compositions with heavy REE (HREE) showing a wide range of values as shown in the spider plots (Fig. 9). Samples of this period returned $(\text{La}/\text{Yb})_{\text{cn}}$ ratios between 3.6 and 117 with a mean value at 47.0, $(\text{Gd}/\text{Yb})_{\text{cn}}$ comprised between 0.97 and 7.34 with a mean value of 3.27, Eu/Eu^* comprised between 0.23 and 1.5 with a mean value at 0.84, Sr/Y values with values from 7.1 and 165 with an average value of 31.5 (Fig. 10).

The late-orogenic stage (2095–2040 Ma) shows also a large spread of composition with HREE spanning over two orders of magnitude. When compared to the previous period, $(\text{La}/\text{Yb})_{\text{cn}}$ ratios overall, continue to decrease and show values comprised between 12.6 and 106.1 with a mean value of 35.0. The $(\text{Gd}/\text{Yb})_{\text{cn}}$ values are comprised between 1.48 and 8.24 with a mean value of 3.3, Eu/Eu^* values are comprised between 0.26 and 1.2 with a mean value at 0.8 and Sr/Y values ranged from 3.3 and 227 with a mean value at 53.9. The overall trace element composition evolution of igneous felsic suites dated in the sWAC with time is shown in Fig. 9 and is summarized in Table 2.

4.3. Lu-Hf data

Lu-Hf analyses were performed on the 49 newly dated samples and on 12 previously dated samples (See Table 1 and ESM3 for details). We first investigate the evolution of the Lu-Hf isotopic signature with time using the same time slices that were defined using the tectonostratigraphic evolution of the craton. We then evaluate the dataset spatial variability.

4.3.1. Temporal evolution

Fig. 11 displays, on an E/Hf vs age plot, both the data collected during

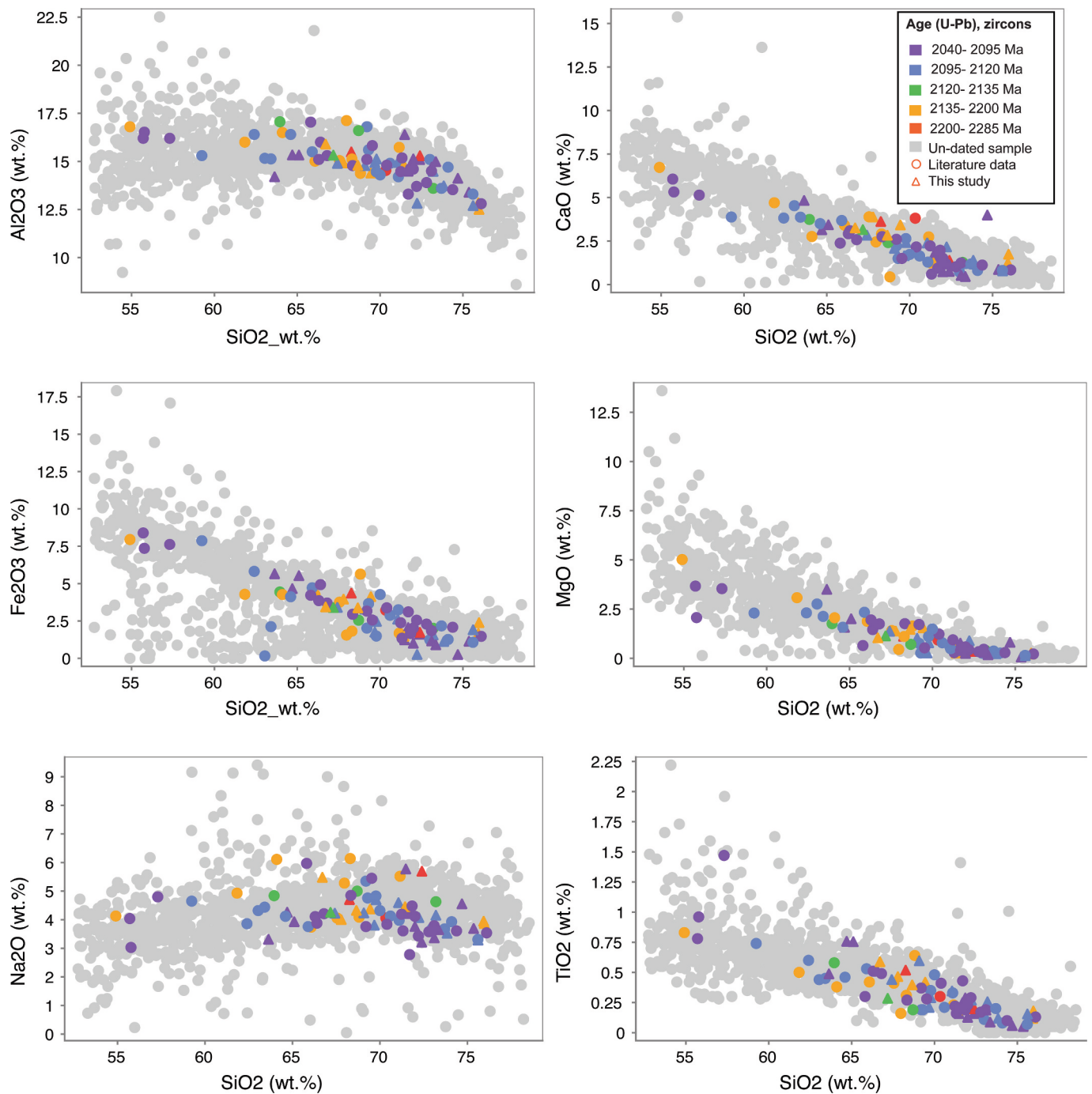


Fig. 5. Selected Harker plots from felsic plutonic and subvolcanic rocks from this study and literature (see ESM4 for references).

this study and the data available in the literature. Such a plot can provide valuable insights into the history of crustal formation, mantle differentiation and reworking processes (Patchett et al., 1981). In this diagram, each data point represents the average ϵ_{Hf} value from ~ 15 zircon grains from each sample with a given magmatic age. Individual Lu-Hf analyses produce a range of values interpreted to reflect mixing between the original melt source and crustal (unradiogenic) material. If the juvenile melt extracted is contaminated by ancient crustal material, then the measured isotopic range may be large, representing in part the initial melt (associated with positive ϵ_{Hf} values if juvenile) and ancient crustal material (negative ϵ_{Hf} values) (Mole et al., 2012). In both cases, the ‘whole-rock’ representative Lu-Hf isotopic composition of the final fluid/magma and, subsequently, the crystallized rock composition,

record a mixture of both sources or composition endmembers.

The five Archean samples included in this study were measured at the margin of the Leo-Man shield in Cote d’Ivoire and span in ages from c. 3400 Ma and 2700 Ma. When combined with the data presented in the literature for this period, ϵ_{Hf} measurements acquired a wide range between -9.43 and $+3.74$, suggestive of both juvenile and recycle crustal input. Associated averaged model ages TDM2 are estimated to be between 3613 Ma and 3064 Ma. The pre-orogenic samples (2285–2200 Ma) present ϵ_{Hf} measurements between $+2.1$ and $+4.9$. Associated model ages TDM2 are estimated to be comprised between 2372 Ma and 2637 Ma. ϵ_{Hf} recorded for early-Eoeburnean plutons (2200–2135 Ma) increase from the previous period and exhibit values comprised between -3.2 and $+7.5$ with a mean value of $+4.2$. Model age TDM2 calculated

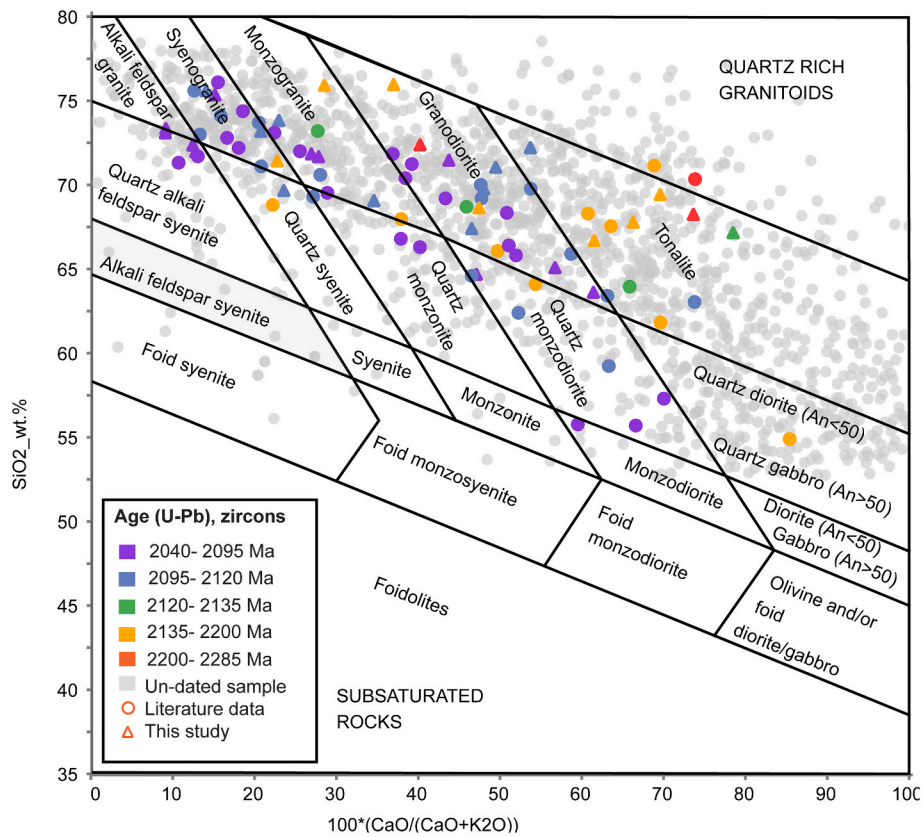


Fig. 6. Classification of plutonic and subvolcanic rocks from this study and literature (see ESM4 for references) after Enrique and Esteve (2019).

for this ϵHf values bracket returned ages comprised between 2145 Ma and 2793 Ma. During the Eoeburnean stage (2135–2120 Ma) ϵHf values are between -2.6 and $+6.9$ with a mean value of $+3.6$. The associated TDM2 model ages that can be calculated from such dataset are comprised between 2193 Ma and 2750 Ma. In the Eburnean stage (2120–2095 Ma), ϵHf values are comprised between -16.3 and $+8.6$ with a mean value of $+2.0$. Model age TDM2 calculated for these ϵHf values bracket returned ages of between 2055 Ma and 3572 Ma. Finally, the late-orogenic stage (2095–2040 Ma) is associated with overall higher ϵHf values comprised between -15.7 and $+8.2$ with mean value of $+3.0$. The associated model ages TDM2 calculated for these ϵHf values are between 3494 Ma and 2077 Ma.

4.3.2. Spatial evolution

This study reconstructs the 4D evolution of the sWAC through the integration of spatially and temporally constrained Lu–Hf isotopic data. The resulting data are visualized as time-slice maps that provide snapshots of the evolving lithospheric architecture during the Rhyacian (Fig. 12).

The first time slice corresponds to the pre-orogenic period from 2285 Ma to 2200 Ma. This period is associated with only four felsic igneous samples emplaced in Burkina Faso, Ghana and Cote d'Ivoire. Their zircons yield a ϵHf signature closer to a chondritic composition than from the depleted mantle estimated at the time. From 2200 Ma to 2135 Ma, the early-Eoeburnean period is associated with a marked increase in igneous activity dominantly over Burkina Faso, Ghana and eastern Cote d'Ivoire but also extending into Mali and Senegal (Kédougou Kénieba inlier). Here the igneous activity is dominated by a juvenile signature with rare negative ϵHf found in southeastern Ghana. From 2135 Ma to 2120 Ma, which corresponds to the Eoeburnean orogeny, ongoing igneous activity continues to affect dominantly the eastern part of the sWAC. Zircons from samples collected in Burkina Faso, Ghana and eastern Cote d'Ivoire continued to exhibit dominantly juvenile signature

with ϵHf values associated with intermediate to near chondritic values. Over the Eburnean orogeny (2120 Ma to 2095 Ma), igneous activity, while still present in Burkina Faso and Ghana, appears to extend further west through Mali and Guinea. Whereas ϵHf values remain dominantly juvenile in Ghana and Burkina Faso, the igneous activity over Cote d'Ivoire, Mali and Guinea appears near chondritic and includes negative ϵHf near the Kénéma-Man Domain. Finally, over the late orogenic stage, igneous activity appears to be concentrated in western Ghana, Cote d'Ivoire, Burkina Faso, Guinea, Mali and Senegal. Over this period most samples yield chondritic or negative ϵHf compositions, with juvenile input that appears to be concentrated in Mali and Senegal.

5. Discussion

5.1. Time space evolution of the sWAC

In the following, we discuss the evolution of felsic igneous activity over the sWAC to frame the crustal growth and its evolution for a period spanning the Archean to the Rhyacian. In Table 3, the tectonic time slices used to frame this tectonic synthesis are summarized with regards to their stratigraphy, plutonism, tectonic, and metallogenic records. The associated summary is modified and expanded for its plutonism component after Perret et al. (2025).

5.1.1. The Archean nuclei (c. 3600 Ma to 2600 Ma)

In the Earth, about half of the internal heat production is provided by the decay of radiogenic elements, whose budget is constrained by geochemical models of the Earth's evolution (e.g. Korenaga, 2018). The other half of the internal heat production derives from the residual heat production including that resulting from the core crystallization. This consensus implies that the Earth was hotter in the Archean and Paleoproterozoic and dissipated more heat than it does today. Such heat dissipation may be achieved by faster spreading ridges (faster mantle

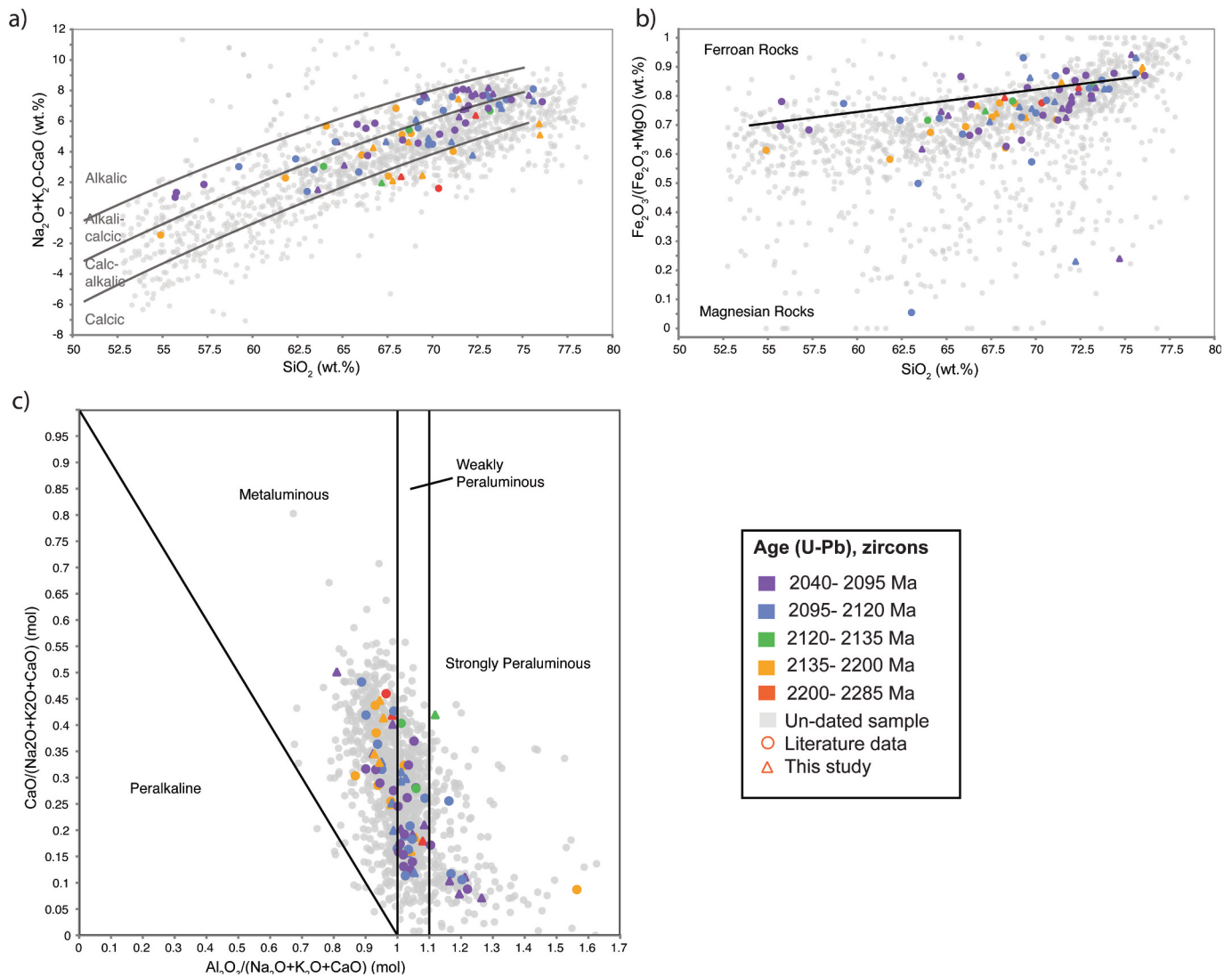


Fig. 7. Plutonic classification diagrams for the samples analyzed both from the literature and in this study (see ESM4). a) modified alkali-lime index (MALI) from Frost et al. (2001), b) $\text{Fe}_2\text{O}_3/(\text{Fe}_2\text{O}_3 + \text{MgO})$ against SiO_2 plot from Frost and Frost (2008), c) alumina saturation plot from Barton and Young (2002) (plotting only samples with $\text{SiO}_2 > 63\%$).

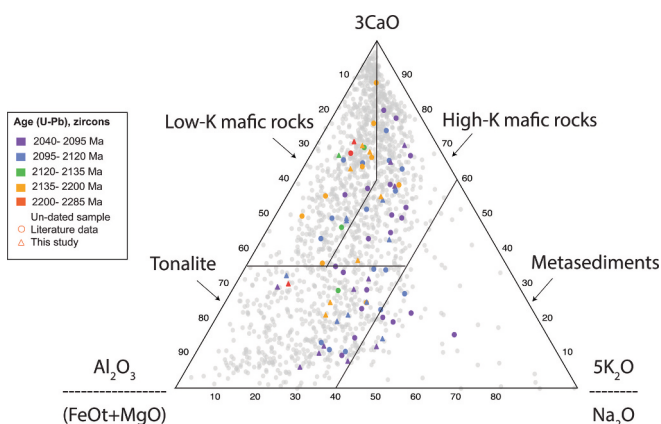


Fig. 8. Tertiary diagram after Laurent et al. (2014) highlighting composition of the source protolith of the felsic igneous rocks presented in this study.

convection) or through having greater ridge length or some combination of both (Scherer et al., 2001; Turcotte and Schubert, 2014). Hargraves

(1986) suggested that heat dissipation is a function of the Earth's surface multiplied by the length of the mid-oceanic ridges. If one assumes that the Earth's surface area remains constant with time, this relationship suggests that with an Archean heat flow three times that of the present-day (Turcotte and Schubert, 2014), twenty-seven times as much ridge length would have been required to account for heat dissipation. This in turn would suggest that the Archean Earth was covered by many small plates moving slowly (Hargraves, 1986). The oceanic basins separating these ancient micro-continental plates were likely consumed through intra-oceanic subduction zones leading to multiple, diachronous collisions between Archean continental blocks of contrasting origin (Liu et al., 2021). Many of these Archean micro-continental plates remain today and are included in the West African craton, i.e., the Reguibat Domain, the Kénéma-Man Domain, the In-Ouzal Domain of the Tuareg Shield and the Archean domains from the Benino-Nigerian Shield (Fig. 13).

However, the Hf data collected on igneous felsic rocks of the Rhyacian age suggests that there were additional Archean domains that were significantly reworked during the Paleoproterozoic such as the Archean BurGhana Domain in southeast Ghana (Block et al., 2016a; Chardon et al., 2020; Petersson et al., 2018, this study) and the Sassandra-Cavally, or SASCA, Domain in southwestern Cote d'Ivoire (Koffi et al.,

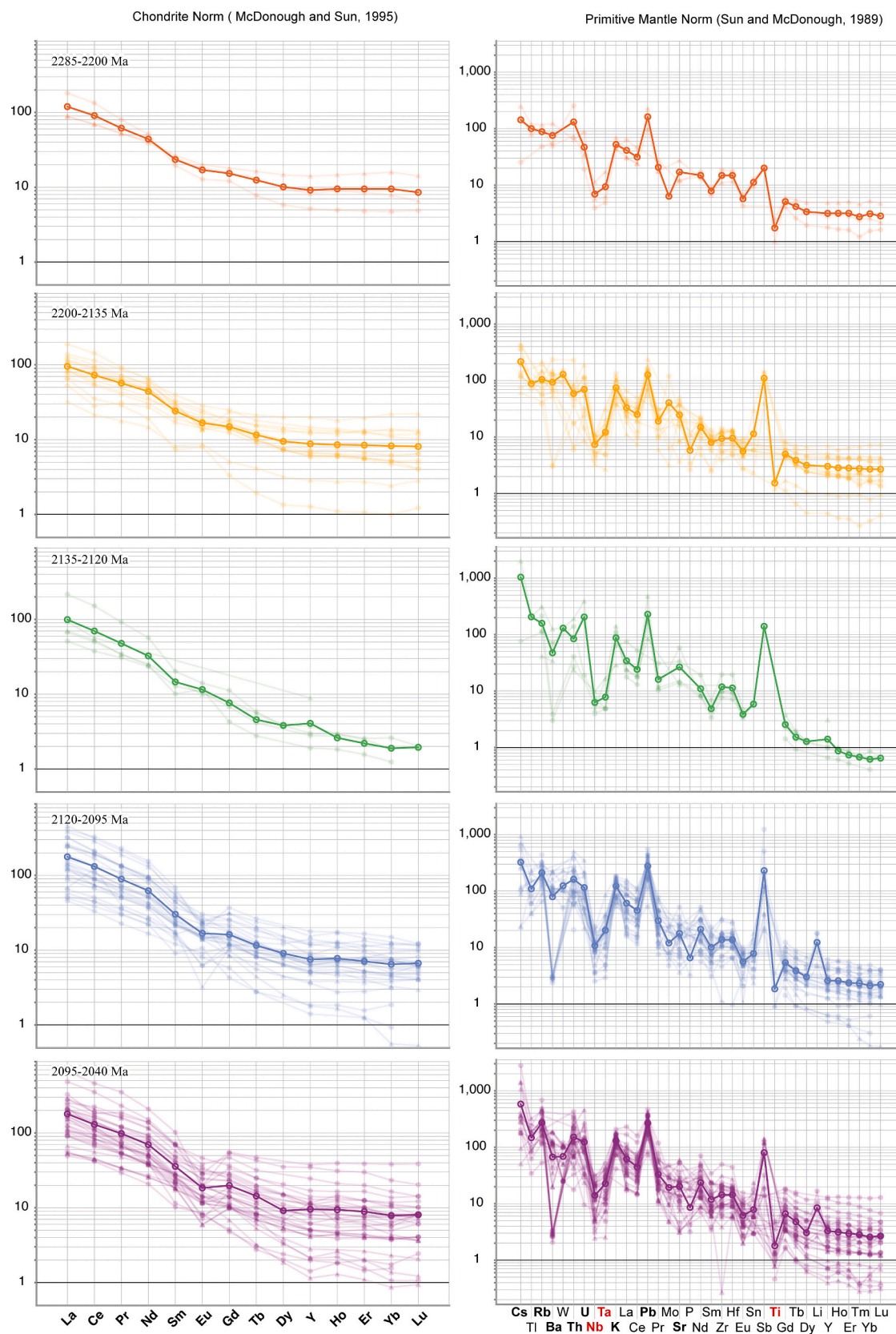


Fig. 9. Spider plots for the felsic igneous rocks that have been dated during this study as well as from the literature (ESM4). Data are presented following reference time periods defined above. Mean values for each element are shown in bold, whereas individual analyses are represented with some transparency.

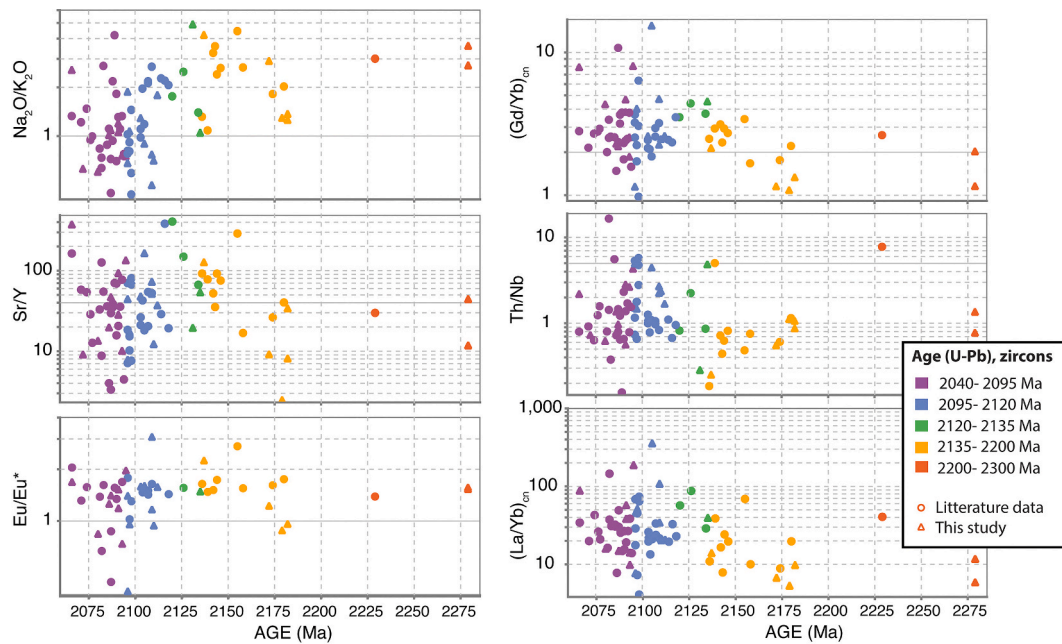


Fig. 10. Selected element ratios plotted against the age showing the geochemical evolution of the igneous felsic rocks composition during the Rhyacian.

Table 2

Summary table of selected major and trace element ratios averaged displaying both range and mean value for each time slice considered.

Geological stage	Age (Ma)	(La/Yb) _n range	Mean (La/Yb) _n	(Gd/Yb) _n range	Mean (Gd/Yb) _n	Eu/Eu* range	Mean Eu/Eu*	Sr/Y range	Mean Sr/Y	Mean Na ₂ O/K ₂ O
Pre-orogenic	2285–2200	5.2–35.9	17.17	1.2–6.6	1.9	0.83–0.94	0.89	11.9–44.6	28.8	3.1
Early Eoeburnean	2200–2135	4.7–57.3	17.2	1.1–3.4	2.2	0.56–1.6	0.96	2.48–204	65.3	2.2
Eoeburnean	2135–2120	35.02–77.15	56.08	3.52–4.55	4.0	0.89–1.76	1.18	19.57–406.67	139	2.0
Eburnean	2120–2095	3.6–117	47.0	0.97–7.34	3.3	0.23–1.5	0.84	7.1–165	31.5	1.2
Late-orogenic	2095–2040	12.56–106.14	35.03	1.48–8.24	3.3	0.26–1.22	0.76	3.34–226.60	53.9	1.1

2022; Kouamélan, 1996). The SASCA Domain has been recently re-interpreted as a strongly reworked portion of the Kénéma-Man Domain, including hybrid, high-grade Paleoproterozoic and Archean crustal fragments (Chardon et al., 2020; Koffi et al., 2022; Traore et al., 2022). With magmatic U-Pb ages acquired on igneous zircon grains spanning c. 3600 to 2600 Ma, these Archean micro-blocks developed over a protracted crustal evolution period spanning nearly 1 Gyr involving both juvenile and crustal reworking stages (e.g. Traore et al., 2022).

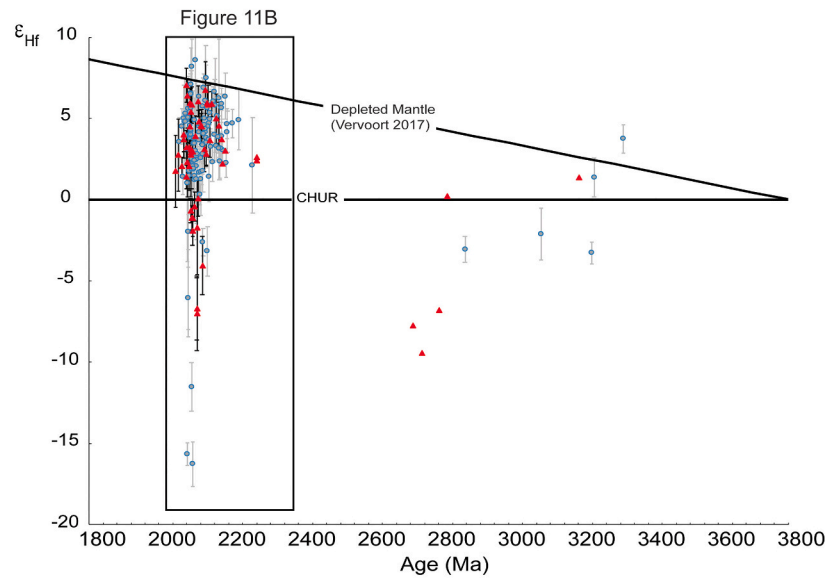
The Archean record of the Kénéma-Man Domain can be tracked back to the Paleoproterozoic with predominant tonalite-trondhjemite-granodiorite gneisses dated between 3500–3600 Ma (Leonian orogeny) in Guinea, western Côte d'Ivoire, Liberia and Sierra Leone (Barth et al., 2002; Bering et al., 1998; Eglinger et al., 2017; Rollinson, 2016; Thiéblemont et al., 2001). A second major igneous event, termed the Liberian orogeny, is recognized over the Kénéma-Man Domain and dated at c. 3260–2850 Ma. The igneous activity recognized is thought to have formed in two episodes between c. 3050–3260 Ma and c. 2850–2960 Ma, where the magmas were derived from the partial reworking of the Paleoproterozoic gneisses (Rollinson, 2016).

Supracrustal cover established over this extended period remains poorly constrained. It largely consists of thick volcanic succession including komatiites, komatiitic basalts and basalts (Rollinson, 1999; Rollinson, 1983) and an upper sedimentary formation comprising greywackes and banded ironstones (Macfarlane, 1981). In Guinea, the Nimba and Simandou supracrustal volcano-sedimentary successions, including metamorphosed mafic igneous rocks, BIFs, and siliciclastic metasedimentary rocks (Thiéblemont et al., 2004), overlie the gneissic

basement, with a maximum deposition age of c. 2615 Ma for the Nimba formation (Billa et al., 1999). The later age marks the cessation of the Neoproterozoic igneous activity in the sWAC.

The Archean Reguibat and the Kénéma-Man Domains are separated by tracts of Paleoproterozoic terrains (e.g. Kédougou-Kénieba Inlier) partially covered by younger sedimentary basins, suggesting that these Archean domains may represent stand-alone blocks. Rollinson (2016), however, compared the Archean evolution of the Reguibat and the Kénéma-Man Domains and noted significant geological similarities, suggesting that these two blocks may once have shared a common history, speculating that they may have been close neighbors or possibly formed part of a single Archean supercraton. Like the Kénéma-Man Domain, the Reguibat Domain, was affected by deformation, granite plutonism, reworking, and high-grade metamorphism during the Liberian orogeny at c. 2800 Ma (Barth et al., 2002; Kouamélan et al., 1997; Rollinson, 2016). The latter may well have marked the final amalgamation and cratonization of various lithospheric blocks such as Reguibat and Kénéma-Man Domains at that time. Several authors further suggested the possible continuation of thinned, reworked Archean lithosphere beneath the Taoudeni Basin based on intermediate wavelength gravity and aeromagnetic anomalies (Begg et al., 2009; Jessell et al., 2016). This echoes with mapped trends of Eburnean faults and shear zones undercover of the Taoudeni Basin, which gradually re-orient in a N-S to NW-SE orientation from the Kénéma-Man Domain towards the Reguibat Domain (Jessell et al., 2016). More recently, Olugboji et al. (2024) employed ambient noise tomography to image crustal velocity structure and identified a discrete domain, termed the “Bou-Naga Domain”, interpreted to represent largely unmodified

A/ ARCHEAN AND RHYACIAN SAMPLES



B/ RHYACIAN SAMPLES

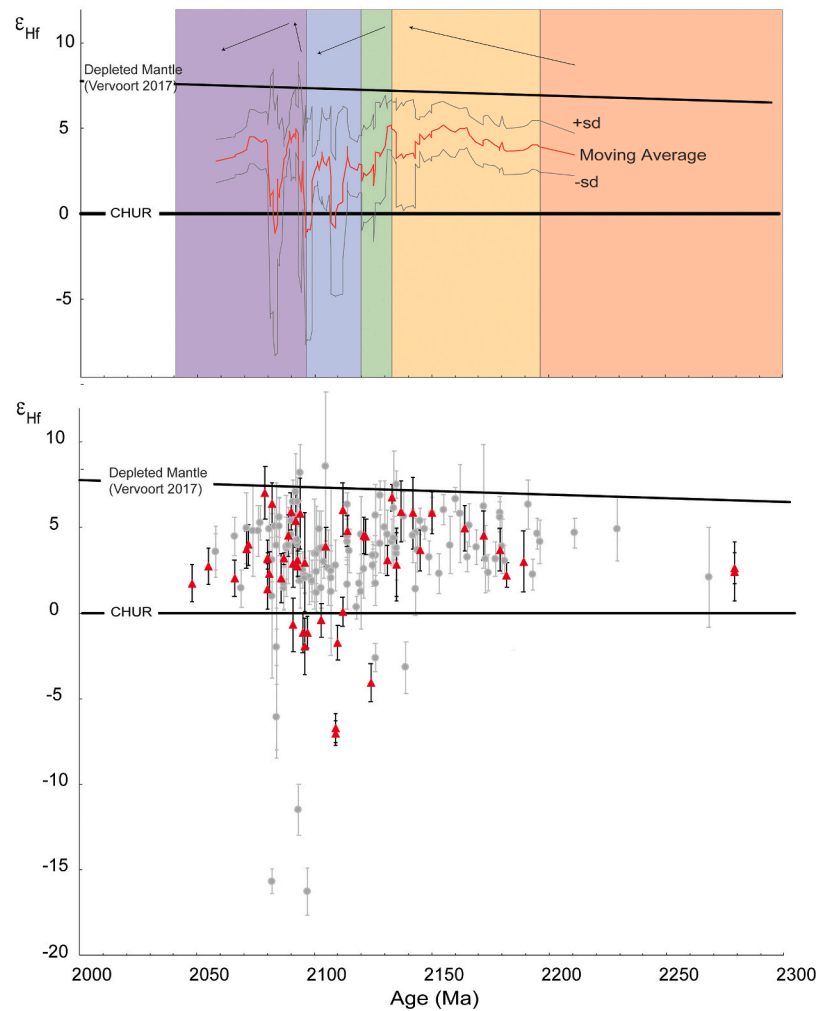


Fig. 11. Epsilon Hf vs age plots for the felsic igneous rocks in the sWAC. The Archean eon and early Paleoproterozoic (4.0 to 2.1 Ga) is characterized by early continental crust formation. During this time, the Earth's mantle was more primitive, and crustal extraction processes were less fractionated compared to later eons. A) data showing the whole dataset including Archean samples red triangle = samples analyzed for this study, grey circles = samples from the literature (Eglinger et al., 2017; Koffi et al., 2022); B) (Bottom) data collected over the Baoulé-Mossi Domain (same legend as for A); (Top) Moving average calculated over 7 data points (dashed line), upper and lower curves are indicative of one standard deviation.

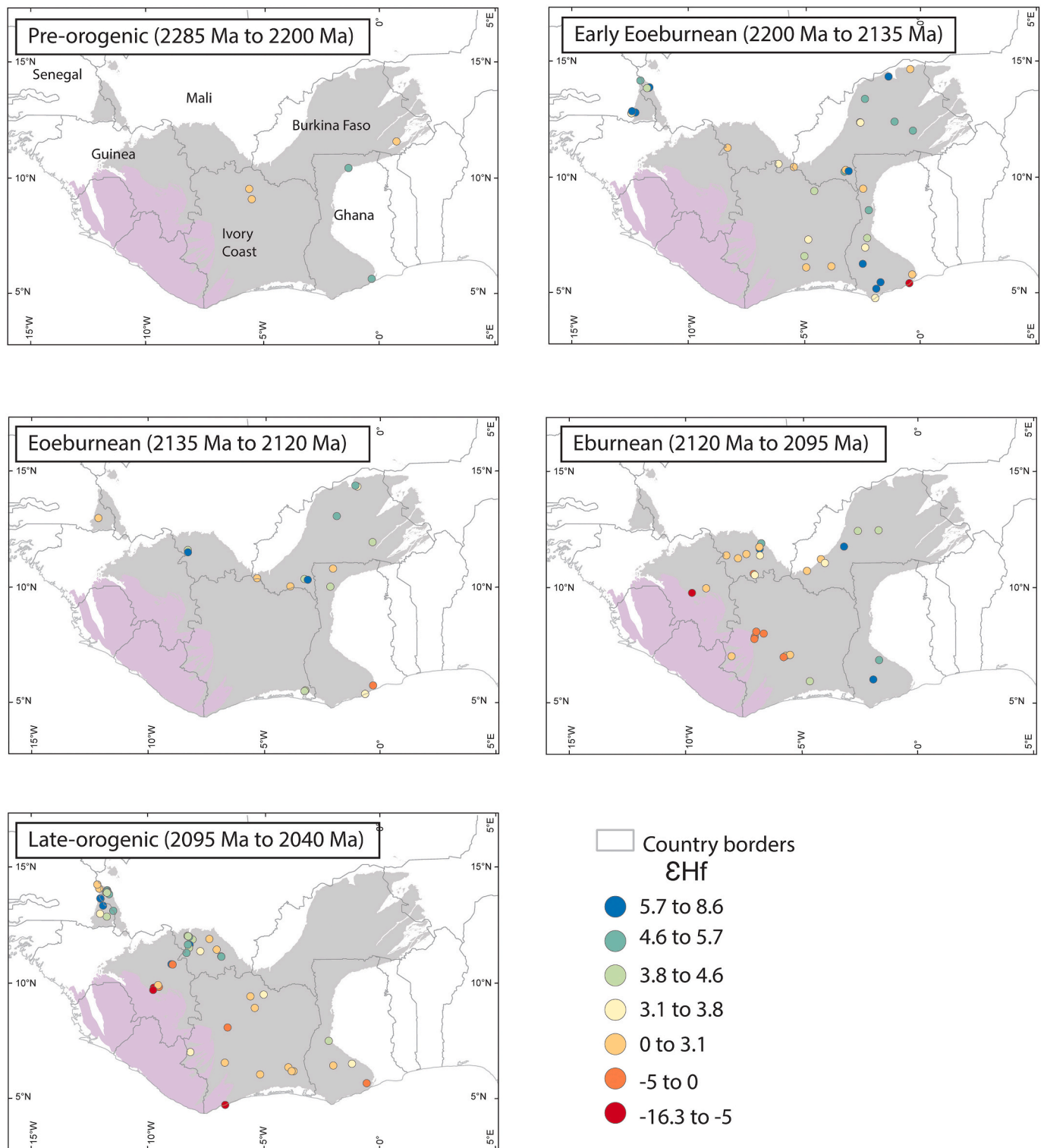


Fig. 12. Time sliced maps showing individual data point with graduated color highlighting the ϵ_{Hf} range of values calculated. The maps are plotted for each swAC Paleoproterozoic period considered in the manuscript. The gray and purple zones represent the location of both the Baoulé-Mossi Domain and Kénéma-Man Domain respectively. See details in the text. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Archean crust beneath the Taoudeni Basin in southern Mauritania and northern Senegal. This domain extends southward beneath basin cover from the Reguibat Domain to the northernmost Kédougou-Kénieba Inlier (Fig. 13). In a similar view, little is known on the potential spatial connection(s) between the BurGhana Domain, the Benino-Nigerian Shield, and the Archean terrane of the Tuareg Shield.

Further away, the Mesoarchean to Neoproterozoic lithosphere in the eastern Samartian region has been documented to exhibit an orogenic record that mirrors that found in Baltica (Terentiev and Santosh, 2020; Wane et al., 2018). The geological similarity observed in these Archean blocks lends substantial support to the hypothesis of a tectonic link between Baltica, the Amazonian Craton, and the West African Craton

Table 3

Time chart summarizing the felsic igneous magmatism, deformation and associated metamorphism, and metallogenic event that punctuated the sWAC from the Archean to the Paleoproterozoic. Modified after [Perret et al. \(2025\)](#).

Stratigraphy	Felsic igneous magmatism	Tectonics: Deformation-Metamorphism	Metallogeny
c. 2095-2040 Ma	Resurgence in bimodal and juvenile mafic-ultramafic volcanism confined to the north-western margin of the Kénéma-Man Domain (e.g. Niandan komatiitic basalts). Volcanism induced by local asthenospheric upwelling and extension triggered by lithospheric delamination or slab break-off	Eburnean tectono-thermal shutdown in the northeastern Baoulé-Mossi Domain at c. 2080 Ma Late Eburnean transcurrent tectonics with sinistral displacement along crustal shear zones in the rest of the Baoulé-Mossi Domain Peak regional metamorphism at 8-10 kbar and 550-600°C in the Baoulé-Mossi Domain and the Kénéma-Man Domain margin, and at 4.5-6 kbar and 500-650°C in the south-eastern Ghana Post-tectonic metamorphism after 2060 Ma: generalised retrogression, and granulite-facies metamorphism and crustal reworking along the eastern margin of the Kénéma-Man Domain (4-14 kbar - 600-1000°C) at c. 2060-2030 Ma, during metacratonisation	Late Eburnean Mineralisation Event: diverse mineralisation styles associated with retrograde P-T-t paths and late collisional tectonics. Late Eburnean Gold Deposits: orogenic and intrusion Au mostly expressed in the western sWAC, newly formed deposits (Bonikro (I), Massawa, Sabodala, Loulo, Sadiola-Yatela) vs overprinting of preexisting mineralisation until c. 2035 Ma (Bonikro (II), Damang (II), Wassa (II), Gaoua (II)) Other Deposits: intrusion-hosted Ni-Cu-PGE (Samapleu), LCT pegmatites (Winneba-Mankoadze, Mangodara, Issia complex) and rare metal granites (Akim Oda district) emplaced until c. 2035 Ma Early Eburnean Mineralisation Event: maximum mineralisation overlapping with the onset of Eburnean deformation, peak metamorphism, and magmatism, particularly in the eastern sWAC (Burkina Faso, Ghana, and Côte d'Ivoire)
c. 2120-2095 Ma	Deposition of shallow to deep water volcanoclastic and epiclastic sedimentary series in the western-central Baoulé-Mossi Domain Ongoing deposition of shallow water to subaerial sedimentary series in narrow depocentres, initiated in the previous time slice	Eburnean Tectonic Event: started around 2120 Ma (up to 2135 Ma), resulted in the collision of the Baoulé-Mossi Domain and the Kénéma-Man Domain around 2095 Ma NW-SE oblique convergence: inversion and folding of volcano-sedimentary series, reverse- to oblique-slip reactivation of basin margin faults and preexisting shear zones Clockwise P-T paths, up to 5-7 kbar and 700-800°C for peak metamorphic conditions in southern Guinea	Early Eburnean Gold Deposits: paleoplacer Au (e.g. Tarkwa, Damang (I)), orogenic Au (e.g. Obuasi (I), Bepkong, Mana (II)), intrusion-related Au (e.g. Morila (I)) Other Deposits: Fe-(Au) skarn (Ity (II)), possible LCT pegmatites (e.g. Goulamina field, related to Morila intrusion?)
c. 2135-2120 Ma	Sedimentary series unconformably overlying earlier volcanic series, local evidence for conformable or disconformable basal contact Peak deposition of shallow to deep water volcanoclastic and epiclastic sedimentary series in ~50-200 km wide depocentres in the eastern-central Baoulé-Mossi Domain Development of slightly younger, narrow depocentres, hosting shallow water to subaerial sedimentary series	Emplacement of syn-tectonic felsic plutons and shallow intrusive rocks likely coeval to voluminous felsic volcanism Spatial distribution and potential chemical heterogeneity poorly documented	No record available
c. 2200-2135 Ma	Peak of the calc-alkaline bimodal volcanism between c. 2200-2175 Ma in the eastern-central Baoulé-Mossi Domain, significant intermediate to felsic magmatism between c. 2175 and 2135 Ma Reworking of felsic volcanic edifices: early deposition of volcanoclastic and epiclastic sedimentary rocks in depocentres, with diminishing volcanic input over time, commenced as early as 2150 Ma Subaerial setting, emerged volcanic centres Cessation of volcanism across the eastern-central Baoulé-Mossi Domain at c. 2135 Ma	Emplacement of calc-alkaline TTG-like plutons Widespread across the Baoulé-Mossi Domain and the Kédougou-Kénieba Inlier Emplacement during crust thickening, from a source dominated by material with a juvenile isotopic signature (mantle-derived?), except for the south-eastern Baoulé-Mossi Domain with evidence for local reworking of Archean crust	Eoeburnean Mineralising Event in the central-eastern part of the Baoulé-Mossi Domain Early Eoeburnean (c. 2200-2170 Ma): VMS (e.g. Perkoa, Tiébébé), porphyry-Cu-(Au-Mo) (e.g. Gaoua (I), Kourki), orogenic Au (e.g. Mana (I), Wassa (I)) Late Eoeburnean (c. 2140 Ma): Au skarn (e.g. Tongon)
c. 2285-2200 Ma	Emplacement of volcanic series overlying older tholeiitic units, with local basal unconformities Transitional to calc-alkaline bimodal volcanism, comprising basalt-andesite-dacite-rhyolite series associated with pyroclastic deposits and co-magmatic gabbro-pyroxenite	Calc-alkaline felsic plutonism associated with the transitional to calc-alkaline bimodal volcanism Mostly reported in the eastern Baoulé-Mossi Domain, but some occurrences documented in the Kédougou-Kénieba Inlier	Identifying mineralisation older than c. 2200 Ma across the sWAC is challenging due to intense subsequent tectono-thermal overprints, limited robust geochronological constraints on multi-commodity ores, and the low preservation potential of certain mineral systems

(continued on next page)

Table 3 (continued)

	Stratigraphy	Felsic igneous magmatism	Tectonics: Deformation-Metamorphism	Metallogeny
	complexes Subaqueous setting to emergence of volcanic centres Onset of calc-alkaline bimodal volcanism likely diachronous between the eastern and western Baoulé-Mossi Domain Emplacement of the basal part of greenstone belts Voluminous mafic volcanism expressed by tholeiitic massive and pillow basalts Subaqueous setting No record available in the WAC.	Plutonism becomes more voluminous after 2200 Ma	and foliated xenoliths in Ghana and north-eastern Burkina Faso	
c. 2350-2285 Ma		No record available	No record available	No record available
c. 2600-23500 Ma	Tentatively, detrital record in Samartia (Baltica) suggest deposition of banded iron formations and terrigenous detrital deposits at c. 2.60–2.45 Ga ago.	Felsic igneous activity associated with syntectonic foliated granite emplacement at 2480 Ma involving older crustal component	Limited record of localised syn-shearing granitic emplacement and granulite metamorphism c. 2480 Ma in the Reguibat.	No record available
c. 3600-2600 Ma	Emplacement of greenstone sequences in the LMD @ c.3600 Ma and 3200 Ma but limited age record available.	TTG emplacement between c. 3600 to 3500Ma. Major granitic magmatism reworking (Leonian) Paleoarchean gneissic basement from 3260 to 2850 Ma (Liberian).	Liberian Orogeny at c. 2800Ma marking the amalgamation of Reguibat Kenema-Man, and Samartia. Greenstone sequences infolded into the TTG gneisses and metamorphosed to amphibolite and granulite grade. No documentation of Mesoarchean tectonic record	Orogenic gold mineralisation documented in Liberia, Sierra Leone and Guinea. Iron ore deposition (Semandou and Nimba formation)

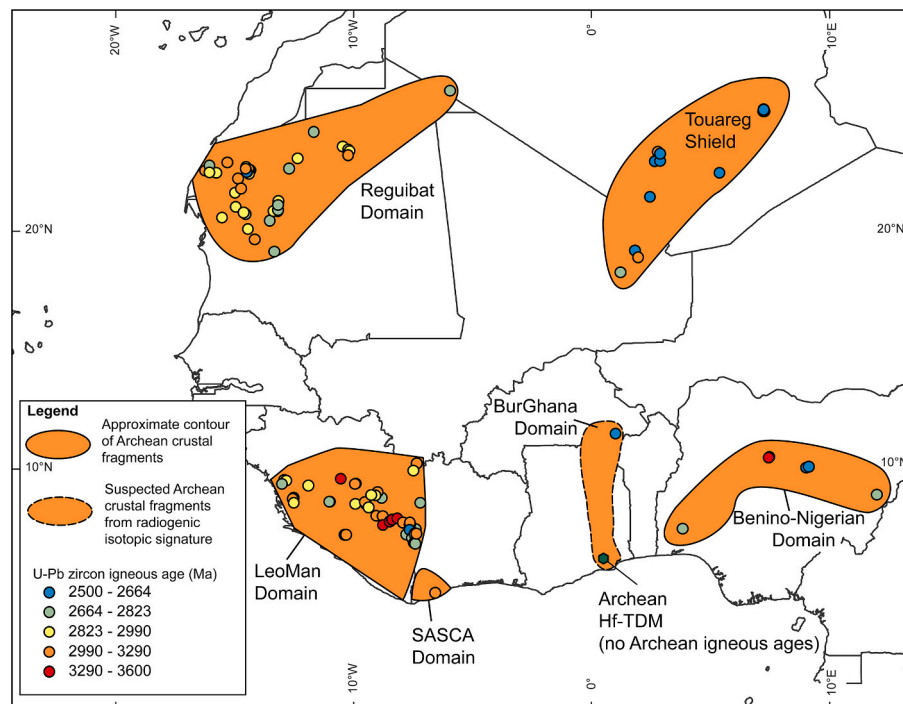


Fig. 13. Map of the West African craton showing the age distribution for U-Pb samples returning Archean age (magmatic or metamorphic). Orange polygons indicate the approximate present-day extent of Archean crustal domains (see ESM5). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

during the Neoproterozoic to Proterozoic era (D'Agrella-Filho et al., 2021; Grenholm et al., 2019a; Johansson, 2009; Reis et al., 2013; Terentiev and Santosh, 2020).

5.1.2. Late Neoproterozoic to the dawn of the paleoproterozoic (c. 2600 to 2350 Ma)

The period following the cratonization of these early Archean micro-

continents is associated with the Siderian period (2500-2300 Ma). From c. 2600 Ma, the WAC is marked by the cessation of active rifting likely associated with a hiatus in both magmatic activity and sediment deposition, which is consistent with regional emersion and peneplain development. Notably, while no sedimentation has been documented to have taken place within the WAC over that period, the rock record in Samartia, which is a Precambrian segment of the East European Craton

of Baltica located in the southern part of eastern Europe, is characterized by the deposition of extensive strata, including banded iron formations and terrigenous detrital deposits at c. 2600–2450 Ma ago (Bibikova et al., 2015; Postnikov, 2002; Savko et al., 2018). Collectively, the partially preserved detrital stratigraphic record that developed during the Siderian is indicative of a passive margin environment, which may tentatively be extended to that prevailing in the West African Craton at the same time.

In the Reguibat Shield, there is evidence for only limited tectonic activity associated with younger regional metamorphism and intrusion of syn-tectonic granites located within major shear zones between c. 2560 and 2480 Ma (Schofield et al., 2012). Sm–Nd depleted mantle model ages indicate that magmatism involved reworking of c. 3200 Ma crust (Schofield et al., 2012). Aside from these localized tectonically active regions, the absence of igneous zircon dated between c. 2600 and 2350 Ma attests to a period of relative magmatic quiescence which lasted until the resurgence of magmatic activity at c. 2350 Ma (e.g. Hayman et al., 2023).

5.1.3. Oceanic basin formation (c. 2350 to 2280 Ma)

The period from c. 2350 to 2280 Ma is dominated by the emplacement of mafic effusive magmas across the Baoulé-Mossi Domain (Abouchami et al., 1990; Hayman et al., 2023; Labou et al., 2020). Basalts and co-magmatic mafic intrusions constitute the base of the lithostratigraphic succession, which has only been robustly dated at 2343 ± 2 Ma in the Oumé-Fettekro greenstone Belt in central Côte d'Ivoire (Hayman et al., 2023). This lowermost mafic lithostratigraphic horizon is characterized by a tholeiitic affinity, relatively flat to slightly depleted REE patterns with low $(\text{La}/\text{Yb})_n$ ratios, and no pronounced HREE depletion. These features are collectively interpreted to be consistent with these primitive melts being derived through partial melting in the spinel-lherzolite stability field (Abouchami et al., 1990; Augustin et al., 2017; Boher et al., 1992; Labou et al., 2020; Lompo, 2009; Ngom et al., 2010; Pouclet et al., 2006a; Sylvester and Attoh, 1992). Further isotopic investigation conducted on both mafic and ultramafic suites interpreted to occupy equivalent stratigraphic levels within the Mako belt in Senegal concluded that the onset of mafic igneous activity associated with a MORB-like composition developed in an embryonic intra-oceanic island arc context and evolved towards a mature intra-oceanic island arc composition (Labou et al., 2020). Investigating both Sm–Nd systematics and trace element analyses of mafic suites deriving from large segments of the WAC, including Mauritania, Senegal, Côte d'Ivoire, Burkina Faso and Niger, suggests that the overall primitive composition, while reflecting MORB-like composition, could also be interpreted as the product of intraplate ocean flood basalts found in Phanerozoic oceanic plateaus (Abouchami et al., 1990). While not mutually exclusive, both interpretations support the development of an oceanic basin opening during the early Eoeburnean, which echoes with the hypothesized passive margins outlined previously.

5.1.4. Pre-orogenic crustal growth (c. 2280 to 2200 Ma)

From around 2280 Ma, the initiation of tonalite-granodiorite-diorite plutonic activity in the eastern sector of the southern West African Craton (sWAC) in Côte d'Ivoire is substantiated by geochronological data from two significant samples, CDIF098 and CDI102A, which both returned a crystallization age at 2279 ± 4 Ma (Table 1). Both igneous crystallization ages extend the onset of felsic magmatism by ~15 Ma when compared with the previous igneous dates of c. 2265 Ma documented in Burkina Faso and Ghana (Parra-Avila et al., 2017; Tshibubudze et al., 2015; Tshibubudze et al., 2013). Cryptic in its record, this early onset of felsic plutonism is coeval with the emergence of both felsic and intermediate effusive volcanic centers (now preserved in the form of orthogneiss), which became widespread after c. 2200 Ma (perhaps as early as c. 2240 Ma). Associated with calc-alkaline compositions, these suites have been shown to accompany the cessation of mafic tholeiitic volcanism in Senegal, Mali, Côte d'Ivoire, and Burkina

Faso (Hayman et al., 2023; Labou et al., 2020; Masurel et al., 2022). In a recent detailed volcanological and geochemical investigation conducted in the Oumé-Fettekro greenstone Belt of central Côte d'Ivoire, Hayman et al. (2023) reported that such bimodal calc-alkaline successions were best interpreted as the result of a nascent intra-oceanic island arc in constructional stage.

Felsic igneous activity during c. 2280–2200 Ma is characterized by the emplacement of voluminous biotite-hornblende-bearing diorite, granodiorite, and tonalite plutonic bodies, predominantly in the eastern part of the Baoulé-Mossi Domain, encompassing Côte d'Ivoire, Burkina Faso, and Ghana (de Kock et al., 2011; Grenholm et al., 2019a; Parra-Avila et al., 2017; Petersson et al., 2016; Tapsoba et al., 2013). Further West in Mali, only a few intrusions have been dated to c. 2230 Ma and 2215 Ma (Lahondere et al., 2002; Parra-Avila et al., 2017) and c. 2200 Ma in Senegal (Gueye et al., 2007), indicating that while the activity was concentrated in the eastern part of the sWAC, fragmental evidence of plutonic activity as old as c. 2200 Ma can be detected across the whole sWAC. The lack of epiclastic rock successions over that early period implies little to no land emersion with the bulk of the Baoulé-Mossi Domain remaining under the sea level.

The felsic igneous rocks from this period exhibit distinct variations in their geochemical and isotopic characteristics, including calcic, metaluminous to slightly peraluminous, and magnesian compositions (Fig. 7). The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios, which average around 0.32, suggest that these felsic igneous rocks possess a sodic composition, likely derived from the partial melting of a primitive mafic crust ($\text{K}_2\text{O}/\text{Na}_2\text{O} < 0.7$ for silicic plutonic rocks; Moyen (2011)). Trace element compositions provide further insights into the source regions of these melts. Among trace elements, the Sr/Y ratio can be used to evaluate the depth of melting, with high Sr/Y values (typically $> \sim 40$) commonly interpreted to reflect deep melting conditions (Lee et al., 2007; Tulloch and Kimbrough, 2003). This signature results from the contrasting stability of plagioclase, which liberates Sr into the melt at relatively high pressures, versus garnet, which retains Y and is stable at greater depths. Although the pressure influence on Sr and Y composition is well documented such ratio can also be strongly influenced by the source composition and possible interaction with mantle-derived melts (Moyen, 2009). The mean Sr/Y and $(\text{Gd}/\text{Yb})_{\text{cn}}$ ratios of 28.8 and 1.9, respectively, indicate that the melt originated from a source region where hornblende and/or garnet were not stable in the cumulate or residue phase (Lee et al., 2007; Tulloch and Kimbrough, 2003). To further evaluate REE patterns, the Eu/Eu^* may help characterize the petromagmatic signature of the source region (Rollinson, 1993). Typically, a negative Eu anomaly indicates that plagioclase was fractionated from these melts during magmatic differentiation, whereas a positive anomaly may reflect plagioclase accumulation and/or suppression of plagioclase crystallization due to the stability of hornblende and/or garnet. Here the Eu/Eu^* ratio indicates a pronounced positive Eu anomaly, suggesting the reworking of a juvenile, low-K mafic source with possible crustal contamination at low to moderate depth.

From an isotopic perspective, the few samples analyzed from this period exhibit a mean EHf value of +3.4, with associated two-stage model ages (TDM2) estimated to range between 2375 Ma and 2576 Ma. These results highlight the juvenile sources to these early melts, consistent with the maturation of juvenile magma extracted from the depleted mantle around 2350 Ma (Hayman et al., 2023). However, the EHf values (+2) recorded for some of these oldest samples suggest some contribution from pre-existing Archean crustal material.

5.1.5. The early Eoeburnean tectonic event (c. 2200 to 2135 Ma)

This period is associated with the onset of the first major deformation event recorded in the sWAC with strain fabrics best preserved in the eastern and northeastern sectors and regrouped under the Eoeburnean orogeny (Block et al., 2016b; de Kock et al., 2011; Hein, 2010; McCuaig et al., 2016; McFarlane et al., 2019; Naba et al., 2004; Perrouy et al., 2012; Tshibubudze et al., 2015; Vanin et al., 2004; Vegas et al.,

2008). Deformation is interpreted to have formed under NNW-SSE-directed bulk crustal horizontal shortening (Block et al., 2016b; Perrouy et al., 2012; Tshibubudze et al., 2015). This period of crustal thickening was associated with the emergence of continental landmasses and reworking of felsic volcanic edifices, which ultimately led to the deposition of volcanoclastic and epiclastic rocks in large-scale sedimentary basins across the eastern Baoulé-Mossi Domain such as the Sunyani, Kumasi, and Maluwe basins (Grenholm et al., 2019a; Masurel et al., 2022; Perret et al., 2025).

Concurrent with this stage, felsic plutonism was also predominantly concentrated in Ghana, Burkina Faso, and eastern Côte d'Ivoire but also punctually extended into Mali and Senegal within the Kédougou-Kéniéba Inlier (Fig. 4). The plutonic rocks from this period continue to display metaluminous to slightly peraluminous and magnesian compositions, with the MALI index indicating an evolution towards more calcic to calc-alkalic compositions (Fig. 7). The K_2O/Na_2O ratios yield a mean value of 0.45, which, as with earlier felsic igneous rocks emplaced during the early Eoeburnean event, suggests a sodic composition derived from the partial melting of primitive mafic crust (Fig. 8). Trace element compositions, with Sr/Y and $(Gd/Yb)_{cn}$ mean values of 65.3 and 2.2, respectively, indicate that the melt was sourced from a region where hornblende and/or garnet were stable in the cumulate (or restite). This composition further attests to the ongoing crustal thickening at that time. U-Pb-Hf isotopic data reveal an increase in ϵHf values, with a mean of +4.2, indicating the generation of predominantly juvenile material in central and western Ghana, Côte d'Ivoire, and Burkina Faso. This juvenile signature contrasts with the negative ϵHf values recorded in granites from the southeastern Baoulé-Mossi Domain in both Ghana and Togo, which suggest the reworking of Archean crust (e.g. Amponsah et al. (2023); Petersson et al. (2018); Petersson et al. (2016)).

5.1.6. Eoeburnean tectonic event (c. 2135 to 2120 Ma)

This period accounts for the deposition of the bulk of shallow- to deep-water epiclastic volcanoclastic sedimentary successions in extensive (~50 to over 250 km wide in present-day width, uncorrected for post emplacement crustal shortening) basins across the eastern and central Baoulé-Mossi Domain. These younger basin successions exhibit a range of conformable to unconformable relationships with underlying strata and are well-documented in regional studies (Adadey et al., 2009; Block et al., 2016a; de Kock et al., 2011; Grenholm et al., 2019b; Hirdes et al., 2007; Oberthür et al., 1998). These features illustrate a continuation of the crustal thickening initiated during the Eoeburnean orogeny and the associated emergence of land masses leading to the deposition of detrital material in large sedimentary basins shortly after the orogenic peak (Perret et al., 2025).

Felsic igneous activity during this stage expanded from eastern Burkina Faso through Mali to Senegal. The associated magmatism is characterized by peraluminous and magnesian compositions, with the MALI index indicating that most igneous felsic rocks have a calc-alkaline to alkali-calcic composition (Fig. 7). The K_2O/Na_2O ratios have a mean value of 0.57, ranging from 0.20 to 0.90, and suggesting a mixed composition that involved a primitive sodic and reworked alkaline source (Fig. 9) involving both mafic and crustal sources (Fig. 8). Trace element compositions, such as Sr/Y and $(Gd/Yb)_{cn}$, yield mean values of 139 and 4.0, respectively. These ratios are indicative of a melt derived from a source region where hornblende and/or garnet were stable in the residue, further supporting evidence of ongoing crustal thickening following the Eoeburnean event.

U-Pb-Hf isotopic data reveal a downward trend in ϵHf values spanning both the early Eoeburnean and late Eburnean periods (see Fig. 11). This trend is consistent with continued crustal thickening and the reworking of early Eoeburnean juvenile suites. Notably, very negative ϵHf values, reaching as low as -4, observed in eastern Ghana and along the eastern and northern margins of the Kénéma-Man Domain, indicate the reworking of Mesoarchean to Paleoarchean crust (Eglinger et al., 2017; Koffi et al., 2023).

5.1.7. Eburnean tectonic event (c. 2120 to 2095 Ma)

Between c. 2120 Ma and 2095 Ma, the locus of sedimentation switched from the east-central Baoulé-Mossi Domain to its northwestern part, with deposition of thick turbiditic series (Allibone et al., 2020; Kone et al., 2020; Lebrun et al., 2016). It is worth mentioning that shallow-water sedimentary successions in the northwestern part of the sWAC include significant volumes of metamorphosed carbonate rocks, a distinctive feature in comparison to eastern counterparts (Grenholm et al., 2019a; Masurel et al., 2022).

In the northwestern part of the sWAC, these sedimentary successions were subsequently followed by the deposition of discordant, shallow-water to subaerial sedimentary series within narrow, fault-bounded, molasse-like sedimentary basins (Adadey et al., 2009; Grenholm et al., 2019b; Hirdes et al., 1996; Kiessling et al., 1997; Lüdtke et al., 1999; Oberthür et al., 1998; Pigois et al., 2003). This detrital sedimentation is closely related to the Eburnean orogeny, during which the predominant structural grain and associated structural fabrics across the Baoulé-Mossi Domain were established. This tectonic event led to the inversion of large-scale sedimentary basins, as well as the development of clockwise pressure-temperature paths in both the eastern and western Baoulé-Mossi Domain (Grenholm et al., 2019a).

In contrast, the Eburnean tectonic event in the northwestern Baoulé-Mossi Domain encompassed the deposition and subsequent inversion of large-scale volcano-sedimentary series (Masurel et al., 2022 and references therein). In this region, the inversion of carbonate-bearing flysch basins between c. 2110 and 2095 Ma resulted in the development of NNE- to NE-trending isoclinal folds, major intra-basin reverse shear zones, and the reactivation of basin margin faults as significant reverse shear zones (Allibone et al., 2020; Allibone et al., 2002a; Allibone et al., 2002b; Hirdes et al., 1992; Masurel et al., 2017; Oliver et al., 2020).

Felsic igneous rocks emplaced during the Eburnean orogeny predominantly exhibit metaluminous, calcic-alkalic to alkali-calcic, and magnesian compositions, with some samples displaying more ferroan characteristics (Fig. 7). In the ternary source plot on Fig. 8, samples appear to be associated with a range of source rocks composition including low-K and high-K mafic composition but also tonalitic and metasedimentary compositions. The K_2O/Na_2O ratios have a mean value of 0.98, indicating a distinctly potassic, evolved crustal source composition (Fig. 9). Trace element compositions, including Sr/Y and $(Gd/Yb)_{cn}$, show lower average values compared to the previous period (31.5 and 3.27 respectively), but they span a wide range of values (Table 2). This variability suggests mixed source regions involving both crustal levels where hornblende and/or garnet were stable and those where plagioclase was stable.

U-Pb-Hf isotopic data for this period continue to show a downward trend in ϵHf values, with both positive and negative values recorded (Fig. 8). This pattern highlights the ongoing reworking of juvenile crust formed during the Eoeburnean and the early Eburnean events, as well as the reworking of Archean crust, particularly around the Kénéma-Man Domain. This isotopic signature reflects the evolution of crustal thickening associated with the Eburnean orogeny and indicates the generation of predominantly reworked continental crust derived from mixed sources across significant portions of the crustal thickness.

5.1.8. Late-orogenic tectonic stage (c. 2095 to 2040 Ma)

The docking between the Archean Kénéma-Man Domain and the Baoulé-Mossi Domain occurred at c. 2095 Ma, following NW-SE-directed oblique convergence (Chardon et al., 2020; Grenholm et al., 2019a; Masurel et al., 2021; Meriaud et al., 2020; Traore et al., 2022). By that time, tectono-thermal activity appears to have largely ceased in the eastern part of the Baoulé-Mossi Domain and shifted entirely toward the northwestern sWAC (Augustin et al., 2017; Grenholm et al., 2019a; Parra-Avila et al., 2017). A resurgence in volcanism is documented to be primarily restricted to the northwestern margin of the Kénéma-Man Domain along the WNW-striking Kankan-Odienné shear zone (Grenholm et al., 2019a; Hirdes et al., 1996; Masurel et al., 2017; Perret

et al., 2025; Thieblemont, 1989), extending to the Kédougou-Kénieba Inlier (Allibone et al., 2020; Hirdes and Davis, 2002; Hirdes et al., 1996; Lahondere et al., 2002; Lambert-Smith et al., 2016).

Bimodal volcanism is associated with the emplacement of both granitic and mafic plutonic bodies. Along the Kénéma-Man Domain in southern Guinea, high-K calc-alkaline to shoshonitic granitic plutonism, dated between 2097 ± 7 and 2082 ± 6 Ma along the Baoulé-Mossi – Kénéma-Man boundary (Eglinger et al., 2017; Lebrun et al., 2016), occurred concurrently with the emplacement of the Samaleu-Yacouba Layered Igneous Complex. This complex, which includes websterites, pyroxenites, and gabbro-norites, intruded along the eastern margin of the Kénéma-Man Domain between 2091 ± 18 and 2080 ± 13 Ma (Gouedji et al., 2014). The bimodal felsic and mafic-ultramafic magmatism between approximately 2095 and 2075 Ma, following a prolonged period of crustal thickening, has been interpreted as the result of asthenospheric upwelling triggered by lithospheric delamination or slab break-off during the aftermath of the collision between the Kénéma-Man and Baoulé-Mossi Domains (Eglinger et al., 2017; Meriaud et al., 2020). This collision has been interpreted as a trigger for the oblique-slip reactivation of several crustal-scale shear zones, across the central and western parts of the Baoulé-Mossi Domain (Egal et al., 2002a; Eglinger et al., 2017; Feybesse et al., 2006).

Felsic igneous rocks emplaced during this late-orogenic stage are characterized by metaluminous to peraluminous, calcic-alkalic to alkali-calcic compositions, displaying both magnesian and ferroan characteristics (Fig. 7). In the ternary source plot on Fig. 8, samples appear to be associated with a range of source rocks composition including dominantly high-K mafic composition as well as tonalitic and metasedimentary compositions. The K_2O/Na_2O ratios have a mean value of 1.04, indicating a clearly potassic, evolved crustal source composition like that observed in samples from the Eburnean stage. Trace element compositions, with Sr/Y and $(Gd/Yb)_{cn}$ mean values of 53.9 and 3.3, respectively, suggest that these rocks were derived from mixed source regions, involving hornblende and/or garnet-stable and plagioclase-stable crustal levels (Fig. 9).

U-Pb-Hf isotopic data for this period show positive ϵHf values around 2095 Ma, indicating the introduction of juvenile material into the crust (Fig. 11). This appears to be coeval with the ongoing reworking of older crustal components around the Kénéma-Man Domain. The presence of both juvenile and reworked signatures is consistent with the bimodal magmatic resurgence involving both crustal and mantle-derived melts. Following this initial peak in ϵHf values, the data show a downward trend highlighting incorporation of Paleoproterozoic continental crust in a post collisional setting (e.g. Wu et al. (2016)).

5.2. Towards a regional geodynamic model for the WAC

The integration of U-Pb-Hf isotopic and whole rock geochemistry data with existing multi-disciplinary datasets encompassing the regional to craton scale offers a unique opportunity to propose a revised and testable geodynamic model for the Paleoproterozoic sWAC (e.g. stratigraphic, structural, metamorphic, and geophysical recently summarized in Perret et al. (2025)). Here we discuss the implications of the sWAC geochemical and isotopic record collected over felsic igneous suites in the context of our collective understanding of the broader West African Craton Rhyacian tectonic and geodynamic configuration.

5.2.1. Multi-disciplinary evidence for Rhyacian convergent margin processes in the sWAC

Plate tectonics on modern Earth are characterized by a kinematically-linked, global interconnected network of internally rigid plates (Dewey, 1969; Harrison, 2020; Korenaga, 2018). In contrast, tectonic modes that potentially operated during the Precambrian remain highly debated (Lenardic, 2018 and references therein). Proposing a coherent tectonic model for a given Precambrian terrain thus fundamentally relies on our ability to analyze and integrate multi-disciplinary

geoscientific data.

The abundance of disparate U-Pb-Hf-zircon evolutionary paths (i.e. ϵHf -time, Fig. 11) in the sWAC can be reconciled with complex tectonic accretion and/or collision between continent/microcontinent and exotic terranes. Similar patterns to those observed in the sWAC have been documented in several Paleoproterozoic provinces worldwide (e.g. Rio Apa Terrane in the Amazonian Craton between c. 2100–1720 Ma, Ribeiro et al., 2020; Minas-Bahia orogenic belt in the São Francisco Craton between c. 2500–2000 Ma, Bruno et al., 2021; North Tarim orogen in the North China Craton between c. 1940–1850 Ma, Ge et al., 2015; Wopmay orogen in northwest Laurentia between c. 2000–1890 Ma, Davis et al., 2015). In addition, cryptic Paleoproterozoic, high-pressure low-temperature eclogite-facies metamorphism (> 10 kbar, $< 450^\circ C$) has been noted in the sWAC (Ganne et al., 2012). We emphasize that the lack of recognition of Paleoproterozoic blueschist-facies series rocks (in contrast to high-pressure granulites and high-temperature eclogites) may be due to the difficulty in deciphering associated pressure-temperature-timing-deformation paths (Baratoux et al., 2024; Ganne et al., 2012) or to poor preservation due to petrological overprinting (e.g. Brown et al., 2022), rather than an actual absence of these rocks in the sWAC (e.g. Palin et al., 2020).

Moreover, the space-time-integrated metallogenic inventory of the Baoulé-Mossi Domain (Masurel et al., 2022; Perret et al., 2025; Thébaud et al., 2020) describes a convergent margin metallogenic cycle, lasting about 100 Myr and best defined as “transitional” between the relatively short-lived (~ 60 –150 Myr), single continental margin metallogenic cycles of late Mesoproterozoic to Neoproterozoic provinces and the longer (~ 220 –440 Myr), multiple-cycle systems typical of Phanerozoic provinces (as defined by Huston et al., 2023). Although there are many limitations and pitfalls associated with the interpretation of whole rock geochemistry data without robust geological context, several features in the space-time-integrated dataset presented here argue in favor of convergent margin processes. The identification of REE patterns in felsic igneous rocks characterized by (i) negative HFSE anomalies, (ii) LILE enrichment and (iii) flat HREE patterns (i.e. garnet-free restite) suggest that parental mafic magmas were either derived during subduction processes (Klemme et al., 2005) or from previously metasomatized sub-continental lithospheric mantle (Hughes et al., 2015). Collectively, the body of data available supports the operation of subduction processes in the sWAC during the Rhyacian. We, however, acknowledge that Proterozoic mantle potential temperatures ~ 80 –150°C hotter than at present likely caused subduction zones to be associated with different geological attributes when compared to their Phanerozoic counterparts (Spencer et al., 2021).

5.2.2. Integrating of the sWAC in the broader West African Craton

Accretionary orogenesis refers to mountain building processes that occur at convergent plate margins where continental crust grows progressively through the addition (accretion) of material derived from subduction-related processes, rather than by a single continent-continent collision (e.g. Cawood et al., 2009). Accretionary orogens, are associated with crustal growth and deformation resulting from some combination of the subduction of oceanic lithosphere and associated arc magmatism (juvenile mantle-derived additions), but may involve the accretion of exotic terranes, including island arcs, oceanic plateaus as well as ancient crustal terranes (Cawood et al., 2009). Within such context, we argue that the Rhyacian tectonic evolution described for the sWAC is consistent with and best defined in terms of accretionary orogenesis involving both Rhyacian and Archean domains. It remains to be evaluated where the main subduction zone(s) and associated suture zone(s) may have been situated within the sWAC. Based on the data presented here, we show that the overall architecture of the sWAC is ultimately controlled by docking of Archean domains against the Baoulé-Mossi Domain: the eastern Baoulé-Mossi Domain collided with the Archean Burghana Domain at c. 2135 Ma (Block et al., 2016a; Grenholm et al., 2019a; Masurel et al., 2022; Petersson et al., 2018)

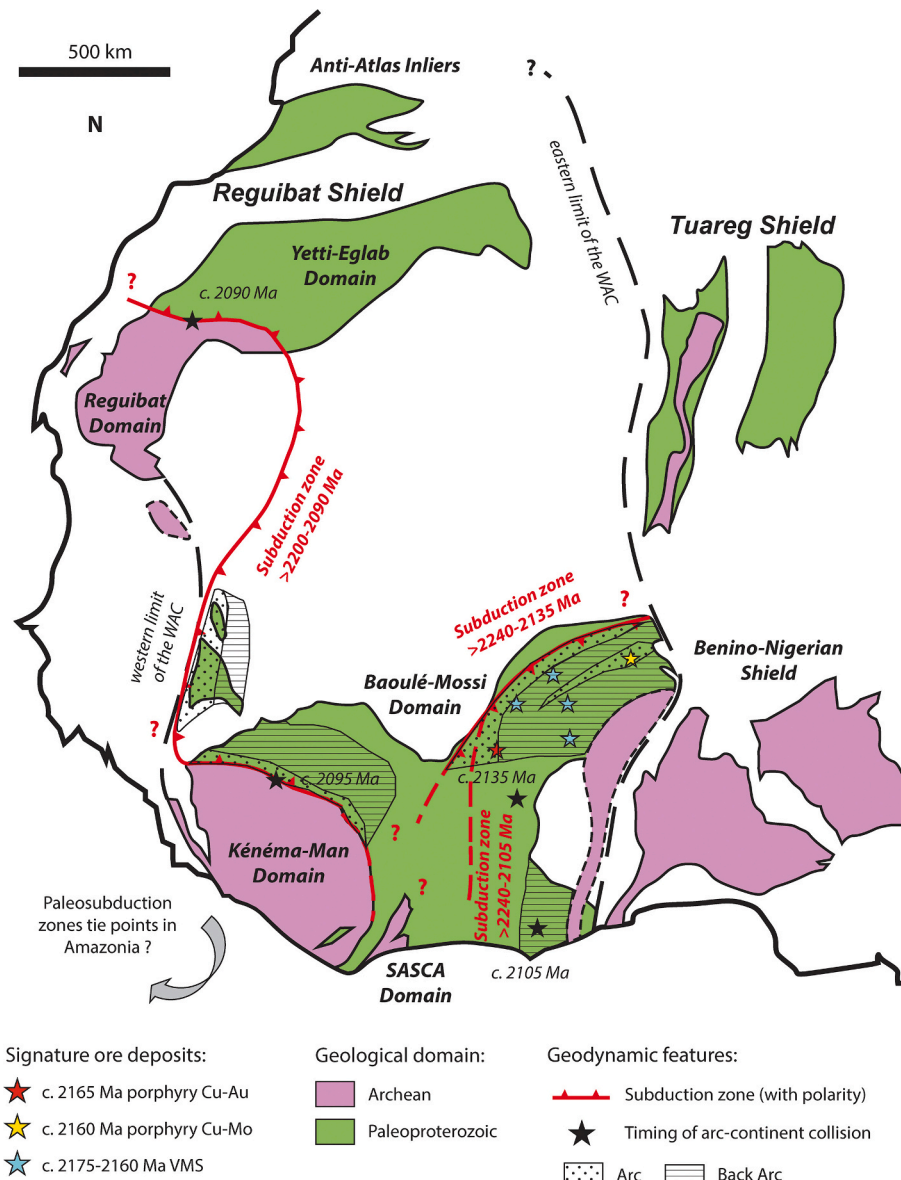


Fig. 14. Tectonic model for the final assembly of the West African Craton.

whereas the western Baoulé-Mossi Domain collided with the Archean Kénéma-Man Domain at c. 2095 Ma (Chardon et al., 2020; Eglinger et al., 2017; Traore et al., 2022). The diachroneity and associated spatial crustal evolution suggests two major segments along which subduction may have occurred (Fig. 12).

5.2.2.1. Eastern segment. The eastern segment is proposed to explain the geological record over Ghana, Burkina Faso and eastern Côte d'Ivoire architecture of the Baoulé-Mossi Domain (Fig. 14). Whereas the exact localization of a subduction suture zone remains ambiguous, the occurrence of several mineral systems (orogenic gold, porphyry Cu and volcanogenic massive sulphides) can be used to estimate its localization and broad geometry (see reviews from Masurel et al., 2022; Perret et al., 2025 and references therein for geochronological constraints). Specifically, the distribution of c. 2175-2155 Ma volcanogenic massive sulphides (Schwartz and Melcher, 2003), porphyry-Cu-Au deposits (Le Mignot et al., 2017) and porphyry-Cu-Mo deposits (Hallarou et al., 2021) may be used as an indirect proxy for the location of a magmatic arc developed in central Côte d'Ivoire and western Burkina Faso along a NE-striking, southeast-dipping subduction zone (Fig. 15). This

hypothesis appears consistent with several key observations, including: (1) intra-oceanic arc magmatism and sedimentation in central Côte d'Ivoire (Hayman et al., 2023; Meriaud et al., 2020; Pouclet et al., 2006b) and southwestern Burkina Faso (Baratoux et al., 2011; Parra-Avila et al., 2019); (2) the significant amount of crustal reworking (zircon ϵ_{Hf} values ranging between 0 to +2, zircon $\delta^{18}\text{O}$ values of 6-10‰ Parra-Avila et al., 2018) that occurred in western Burkina Faso, interpreted to reflect <10% contribution of subducted Archean-aged detritus beneath an interpreted magmatic arc (Parra-Avila et al., 2019); and (3) the Greenville-Ferkessedougou-Bobo-Dioulasso Shear Zone (Fig. 1) that runs from southwestern to north-central Côte d'Ivoire coincides with the most voluminous and strike-extensive, peraluminous and alkaline composite batholith currently recognized across the SWAC (Dombia et al., 1998), possibly reflecting post-collisional partial melting within an accretionary prism between c. 2095-2085 Ma. According to this interpretation, southeastern Ghana may have developed in a back-arc setting compatible with significant input of juvenile material and the production of felsic igneous rocks associated with a garnet stable source region (Fig. 13).

This eastern subduction was active at the latest at c. 2240 Ma and

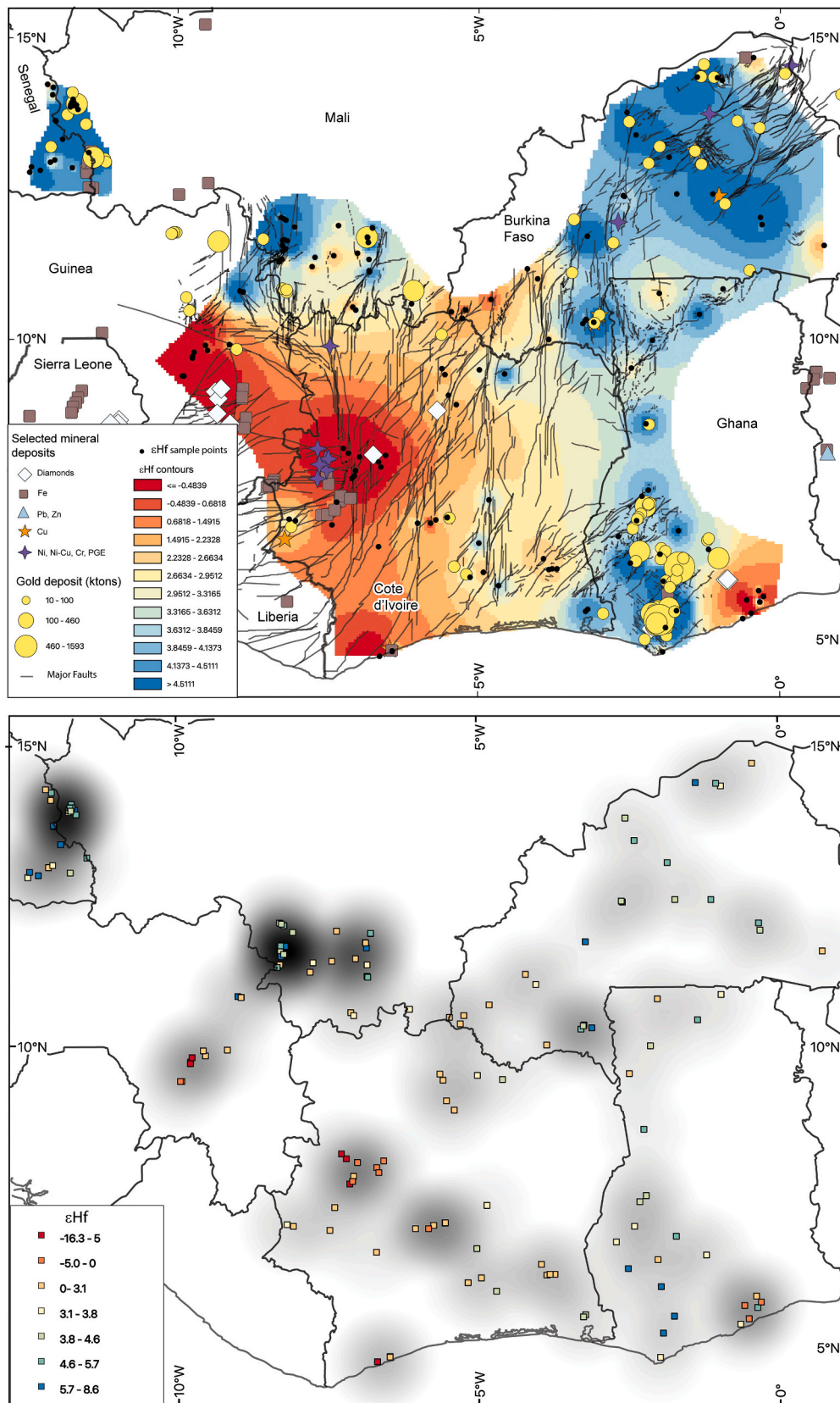


Fig. 15. Interpolation of ϵHf including all data for the period 2300 Ma to 2050 Ma using the IWD algorithm in QGIS (weighted power of 2, isotropic interpolation), A) Epsilon Hf including all data for the period 2300 Ma to 2050 Ma. On this map, mineral deposits (Markwitz et al., 2016) are overlain together with the major structural elements mapped using combination of potential field data and geological datasets, B) Heat map showing sparse data distribution (radius = 35 km averaging dimensions of felsic igneous centres).

terminated at c. 2135 Ma in northeastern Burkina Faso and north-western Ghana (Block et al., 2016a; Grenholm et al., 2019a; Masurel et al., 2022; Petersson et al., 2018). The defined onset of subduction at c. 2240 Ma is correlated with the emergence of voluminous, bimodal, calc-alkaline volcanism and coeval metaluminous to slightly peraluminous, biotite-hornblende-bearing diorite-granodiorite-tonalite plutonic bodies within greenstone belts across the eastern Baoulé-Mossi Domain (this study; de Kock et al., 2011; Grenholm et al., 2019a; Masurel et al., 2022; Parra-Avila et al., 2017; Petersson et al., 2016; Tapsoba et al., 2013). We hypothesize that the timing of oceanic basin closure was diachronous from northeast to southwest, with inversion and closure of back-arc basin depocenters in southeastern Ghana occurring after c. 2120 Ma, nearly 15 Myr later after the initial collision between the northeastern Baoulé-Mossi and BurGhana Domains. This appears consistent with large tracts of the northeastern Baoulé-Mossi Domain reaching lock-up stage against the BurGhana Domain and undergoing transcurrent tectonics after c. 2135 Ma while deposition of voluminous turbiditic successions in large sedimentary depocenters occurred broadly coevally between c. 2135–2120 Ma in the southeasternmost Baoulé-Mossi Domain (Grenholm et al., 2019a; Masurel et al., 2022).

5.2.2.2. Western segment. Deciphering the location and geometry of the other hypothesized subduction zone in the northwestern Baoulé-Mossi Domain is made more difficult as it is likely covered by the Taoudeni Basin and associated with comparatively few robust U-Pb-Hf isotopic constraints. Nevertheless, based on the empirical observation that most giant Precambrian gold provinces worldwide formed in back-arc basin settings (e.g. Groves et al., 2020; Hronsky et al., 2012), one may speculate upon the existence of a southeast-dipping subduction zone associated with a magmatic arc located outboard of the Kédougou-Kénieba Inlier. Such speculation would suggest that the Kédougou-Kénieba Inlier to have formed in an arc and associated back-arc setting which is compatible with its world-class orogenic gold endowment (Allibone et al., 2020; Masurel et al., 2017). This interpretation would also be compatible with the emplacement of highly juvenile felsic igneous rocks in the Kédougou-Kénieba Inlier (Fig. 15).

Whereas additional work is needed to critically evaluate this hypothesis, the inferred geometry and polarity are broadly consistent with observations from the Reguibat Shield in the northern part of the West African Craton. The Reguibat Shield consists of the Paleoproterozoic Yetti-Eglab Domain (Taylor et al., 2019) and the Archean Reguibat Domain (Peucat et al., 2005). Most authors agree that a significant tectono-thermal event occurred between c. 2100–2070 Ma in the Reguibat Domain and marked the collision between the Yetti-Eglab Domain and the Reguibat Domain (Peucat et al., 2005; Schofield et al., 2006; Schofield and Gillespie, 2007; Taylor et al., 2019). Structural studies in the Archean/Paleoproterozoic contact zone are consistent with E-W-directed bulk crustal shortening and sinistral transpression between c. 2090 and 2075 Ma (Kahoui et al., 2008; Merabet et al., 2016; Peucat et al., 2005). Crustal-thickening-related anatexis along this contact zone is constrained to c. 2120–2100 Ma (Combs, 2018; Schofield et al., 2006). Syn-collisional diorite-tonalite-granodiorite plutons of the Chegga Suite in the Eglab Terrane record crystallization ages between 2095 ± 3 Ma and 2091 ± 3 Ma (Peucat et al., 2005). Similarly, a syn-collisional tonalite intruding the Archean-Paleoproterozoic contact zone yielded a crystallization age of 2099 ± 19 Ma (Lehbib, 2016). Syn-orogenic felsic volcanism of the Oued Souss Suite erupted at 2087 ± 9 Ma in the Eglab Terrane, unconformably overlying deformed and metamorphosed basement dated at c. 2210–2180 Ma (Peucat et al., 2005). In contrast, voluminous, undeformed high-K felsic magmatism, including (i) the 2070 ± 3 Ma alkali-calcic Aftout granites, monzogranites, and syenogranites; (ii) ca. 2075–2065 Ma calc-alkaline Eglab rhyodacites and rhyolites; and (iii) the 2076 ± 16 Ma Djebel Drissa alkaline to peralkaline granite, has been interpreted as post-collisional

in origin (Peucat et al., 2005). Consequently, ca. 2070 Ma is taken as a minimum age for the termination of subduction along the western West African Craton.

Based on the above, the emerging view is that c. 2210–2070 Ma geological evolution of the Reguibat Domain can be reconciled with a single, east-dipping subduction zone responsible for accretion and c. 2090 Ma collision of an arc terrane with an Archean block to the West (Taylor et al., 2019). It seems reasonable to link east to north-dipping subduction followed by collision between the Reguibat and Paleoproterozoic Yetti-Eglab Domains at c. 2090 Ma (Combs, 2018; Lehbib, 2016; Peucat et al., 2005; Taylor et al., 2019) with the hypothesized southeast-dipping subduction zone and collision between the northwestern Baoulé-Mossi Domain and Kénéma-Man Domain at c. 2095 Ma. Several authors have also documented similar timing and kinematics in the contact zone of the Reguibat Domain (Combs, 2018; Schofield et al., 2006) compared to that of the Kénéma-Man Domain margin (Chardon et al., 2020; Traore et al., 2022). This idea is given further legitimacy when considering the absolute timing, extent, and compositional spectrum of c. 2090–2070 Ma post-collisional magmatic suites in the Yetti-Eglab Domains and Archean-Paleoproterozoic transition zone of the sWAC, interpreted in both regions to reflect lithospheric delamination or slab break-off (Combs, 2018; Eglinger et al., 2017; Meriaud et al., 2020; Peucat et al., 2005; Taylor et al., 2019). Lastly, it is interesting to note that the Dialé-Daléma and Kofi series in the Kédougou-Kénieba Inlier contain thick beds of marble not observed anywhere else in the Baoulé-Mossi Domain. Coincidentally, the Paleoproterozoic Leareig and Sfariat-Bu Daira terranes (Combs, 2018), correlated with the “contact zone” historically defined in Schofield et al. (2006), also contain up to 1 km-thick marble units, unique to the Reguibat Shield and in a similar location relative to the proposed tectonic arrangement at the time.

Available geochronological constraints on the onset of calc-alkaline magmatism in both the Kédougou-Kénieba Inlier and Yetti-Eglab Domain, taken as a proxy for the onset of intra-oceanic subduction, appear permissive with this model. Indeed, magmatic activity in both the Kédougou-Kénieba Inlier and Yetti-Eglab Domain started at or earlier than c. 2210 Ma (Gueye et al., 2007; Peucat et al., 2005). Based on current knowledge, we hypothesize that the western subduction zone was active at the latest by 2210 Ma and terminated at c. 2090 Ma. The polarity of this east-dipping subduction zone in the Reguibat Domain is based on the passive nature of the contact zone as reflected by the lack of Rhyacian arc-related magmatism at the Archean domain margin (Schofield et al., 2006). Such polarity appears consistent with the lack of magmatism of continental arc affinity recorded around the Kénéma-Man Domain margins prior to c. 2090 Ma (Eglinger et al., 2017; Grenholm et al., 2019a).

5.3. Speculation on the Rhyacian basin geometry

Two tectonic scenarios may be envisaged based on the geological record presented above. In a first interpretation, the overall tectonic record of the sWAC may be interpreted as the result of a strike-extensive (1000 km), intra-oceanic subduction zone, followed by diachronous collision of this intra-oceanic arc system with Archean continental blocks (~2200–2070 Ma) (Grenholm et al., 2019a) akin to that observed today in the Philippines Archipelago (Aurelio et al., 2013). One of the main limitations associated with such an extended and continuous subduction is that it would require a change of subduction polarity between the eastern and western segments. This scenario would require a complex geometry equivalent to what is currently observed in New Zealand with two subduction zones of opposite polarity linked by a lithospheric-scale transfer zone. As an alternative, as both structures show some diachronic activity and yet overlap in time, it is possible to propose a scenario involving two independent but partly contemporaneous intra-oceanic subduction zones.

Several publications infer the amount of oceanic crust in the Paleoproterozoic Baoulé-Mossi Domain to be very limited and largely reflect

strongly attenuated/thinned (Archean) continental crust, possibly hyperextended and/or reworked by mantle-derived magmas (e.g. Petersson et al., 2018; Petersson et al., 2016). We refer to this hypothesis as the confined setting model, whereby a narrow (<500 km wide) oceanic basin formed in between a mosaic of Archean lithospheric blocks. Other workers favor a large (>1000 km wide) oceanic realm model (e.g. Grenholm et al., 2019a; McFarlane et al., 2019). When tested beyond geochemical signatures, the debate fundamentally centers on a space problem, which mainly concerns the model of ~100 Myr of subduction strongly advocated in recent reviews of Paleoproterozoic tectonics in the sWAC (Grenholm et al., 2019a; Perret et al., 2025).

Although the geodynamic modes that prevailed at the dawn of the Proterozoic and the influence of secular changes on the dynamics of (proto-)subduction zones remain debated (Cawood et al., 2022; Hawkesworth et al., 2024; Spencer et al., 2021; Stern, 2023), we have discussed above key arguments why we favor a uniformitarian model for the Paleoproterozoic WAC. In the Phanerozoic eon, subduction rates in intra-oceanic island arc systems range from 2 cm/yr (e.g. Lesser Antilles) to 24 cm/yr (e.g. North Tonga), with typical rates in the range of 5–13 cm/yr (Stern, 2011). Assuming a subduction rate of 3.5 cm/yr, as defined by (Parra-Avila et al., 2017) based on identified time-space-integrated migration of magmatism across the sWAC, Rhyacian subduction would have consumed an ocean on the order of ~3,500 km wide. We concede that this value is not perfect due to unknown spreading rates or total amount of crust recycled along one or two contemporaneous subduction zones. Our calculation also assumes a single ocean (i.e. no back-arc spreading), with a subduction rate higher than the spreading rate. We emphasize, however, that the estimated duration of subduction used here is conservative.

Our calculations hint at the existence of a prevailing oceanic realm between c. 2350 and 2200 Ma, which is further supported by multi-disciplinary lines of evidence including lithofacies associations characteristic of the lower stratigraphic record, geochemical signatures of igneous rocks, Sm-Nd isotopic data on igneous rocks, and U-Pb-Hf igneous zircon systematics and associated space-time-integrated isotopic arrays. Regardless of which scenario is favored, these calculations indicate the existence of a large oceanic basin, contrasting with earlier propositions that associated the Archean period with smaller oceanic plates and slower plate velocities (Hargraves, 1986). This discrepancy may underscore a pivotal transition during the Rhyacian, suggesting that tectonic and underlying geodynamic processes were evolving to resemble what is observed in the Phanerozoic, particularly as secular cooling progressed.

6. Conclusion

Crustal growth in the southern West African Craton unfolded across the late Siderian to the dusk of Rhyacian, from the opening of an oceanic basin at c. 2350 Ma to the aftermath of Archean Kénéma-Man – Paleoproterozoic Baoulé-Mossi domains collision at c. 2040 Ma. Two orogenic cycles, the Eoeburnean (~2240–2135 Ma) and the Eburnean (~2120–2090 Ma), record the progressive closure of this basin through pulses of juvenile magmatism and reworking of Archean crust.

The Eoeburnean orogeny is marked by melts from garnet- and/or hornblende-stable sources, consistent with crustal thickening and dominantly juvenile ϵHf signatures across central and western Ghana, Côte d'Ivoire, and Burkina Faso. Negative ϵHf values in the southeastern Baoulé-Mossi Domain (Ghana, Togo), by contrast, record reworking of Archean crust.

The Eburnean orogeny marks the tectonic peak of the sWAC. Melting spanned both garnet/hornblende-stable and plagioclase-stable crustal levels, and ϵHf values shifted further downward. These observations are a key evidences of continued reworking of Eoeburnean juvenile crust and Archean basement during that period, particularly around the Archean Kénéma-Man Domain.

Late-orogenic bimodal magmatism (c. 2095–2040 Ma) along the

Kénéma-Man domain margin in southern Guinea reflects asthenospheric upwelling following lithospheric delamination or slab break-off. A positive ϵHf pulse at c. 2095 Ma records renewed juvenile input, followed by a final downward trend as Paleoproterozoic crust was incorporated into a convergent setting.

We interpret this record as evidence for two partly synchronous subduction segments: an eastern, southeast-dipping segment (c. 2240–2135 Ma) that drove Eoeburnean collision between the Baoulé-Mossi Domain and the Archean BurGhana block, and a western, east-dipping segment (c. 2210–2070 Ma) that extended north into the Reguibat Shield and drove Eburnean collision with the Kénéma-Man Domain. Whether this reflects one continuous, polarity-reversing subduction system or two independent zones remains open, but either scenario requires closure of an oceanic basin roughly 3,500 km wide.

CRediT author contribution statement

Nicolas Thébaud: Conceptualization, Field sampling, Data curation, Formal analysis (geochronology, isotopic geochemistry, petrography), Writing – original draft. **Quentin Masurel:** Conceptualization, Field sampling, Data curation, Writing – review & editing. **Luis A. Parra-Avila:** Conceptualization, Writing – review & editing. **Aurélien Eglinger:** Conceptualization, Field sampling, Formal analysis (geochronology, isotopic geochemistry, petrography), Writing – review & editing. **Julien Perret:** Visualization, Writing – review & editing. **Augustin Yao Koffi:** Conceptualization, Field sampling, Formal analysis (geochronology), Writing – review & editing. **Ousmane Wane:** Conceptualization, Field sampling, Formal analysis (petrography), Writing – review & editing. **Helen McFarlane:** Conceptualization, Writing – review & editing. **Graham Begg:** Conceptualization, Writing – review & editing. **Guillaume Duclaux:** Conceptualization, Writing – review & editing. **Mark W. Jessell:** Conceptualization, Writing – review & editing. **Olivier Vanderhaeghe:** Conceptualization, Writing – review & editing. **Lenka Baratoux:** Conceptualization, Writing – review & editing. **Mahamadou Diallo:** Writing – review & editing. **Alain N. Kouamelan:** Field logistical support and sampling. **Anthony I.S. Kemp:** Resources (laboratory support), Data processing, Writing – review & editing. **Noreen J. Evans:** Resources (laboratory support). **Armin Zeh:** Resources (laboratory support).

Declaration of competing interest

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.earscirev.2026.105624>.

Data availability

The data used for the conceptualisation of this manuscript have been included in the supplementary materials and made available.

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