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Tightening-up NE Brazil and NW Africa connections: New U–Pb/Lu–Hf zircon data of a complete plate tectonic cycle in the Dahomey belt of the West Gondwana Orogen in Togo and Benin

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ABSTRACT

The Dahomey belt in Togo and Benin is an important segment of the larger West Gondwana Orogen. Here, we review the geodynamic evolution of the Dahomey belt and discuss new U–Pb and Lu–Hf zircon data in light of similar data previously acquired on the geologically related Northern Borborema Province, in NE Brazil. Eighteen samples from different tectonic settings and regions within the belt were collected for zircon isotopic investigation. Passive margin deposits of the Atacora Structural Unit and lower units of the Volta Basin have detrital zircon signatures compatible with the flanking West Africa Craton. The arc-related magmatism resulted from the east-dipping subduction of the Goiás-Pharusian oceanic lithosphere and is represented by a variety of granitoids emplaced in the Benino-Nigerian Shield between 670 and 610 Ma. These granitoids were mainly sourced from crustal reservoirs with subordinate juvenile input. Detrital zircon ages from syn-orogenic deposits in Benino-Nigerian Shield suggest that arc development could have started as early as 780 Ma. The main period of melting in the internal part of the belt, the Benino-Nigerian Shield, is related to crustal thickening and occurs only ca. 30 m.y. after initiation of the continental collision, marked by the ca. 610 Ma ultra-high pressure (UHP) metamorphism recorded at Lato Hills. Foreland development represented by the upper units of the Volta basin developed soon after continental collision and persisted with the development of the west-verging thrust front synchronously with the main period of crustal melting due to collision at ca. 580 Ma. The subvertical Transbrasiliano Lineament in South America, that corresponds to the Kandi Lineament in Africa, provides a present-day fit between NW Africa and NE Brazil. Restoration of the movement of the Transbrasiliano-Kandi Lineament (strike-slip plate boundary) places the Dahomey belt and Borborema Province (NE Brazil) along the same section of the West Gondwana Orogen. This configuration would explain some of the misfits previously discussed in the literature and aligns the UHP eclogites in Togo and NE Brazil.

1. Introduction

Geological correlations between Brazil and Africa have been proposed since the advent of plate tectonics theory (Hurley et al., 1967; Almeida and Black, 1968) or even earlier (see De Wit et al., 2008a,b and references therein). The Neoproterozoic events in Africa (Kennedy, 1964), particularly those recorded along the eastern edge of the West African Craton (WAC) have been grouped in the well-defined Trans-Saharan orogeny (Caby, 1989) and its continuation into South America has been extensively debated (e.g. Caby, 1989; Santos et al., 2008; Arthaud et al., 2008; Cordani et al., 2013a,b; Kalsbeek et al., 2012). Based on the synchronous timing of the continental collision constrained in each sector, this extensive orogenic area was recently re-assembled into the major West Gondwana Orogen (WGO), which extends from present-day northeast Africa to central Brazil (Ganade de Araujo et al., 2014a) (Fig. 1).

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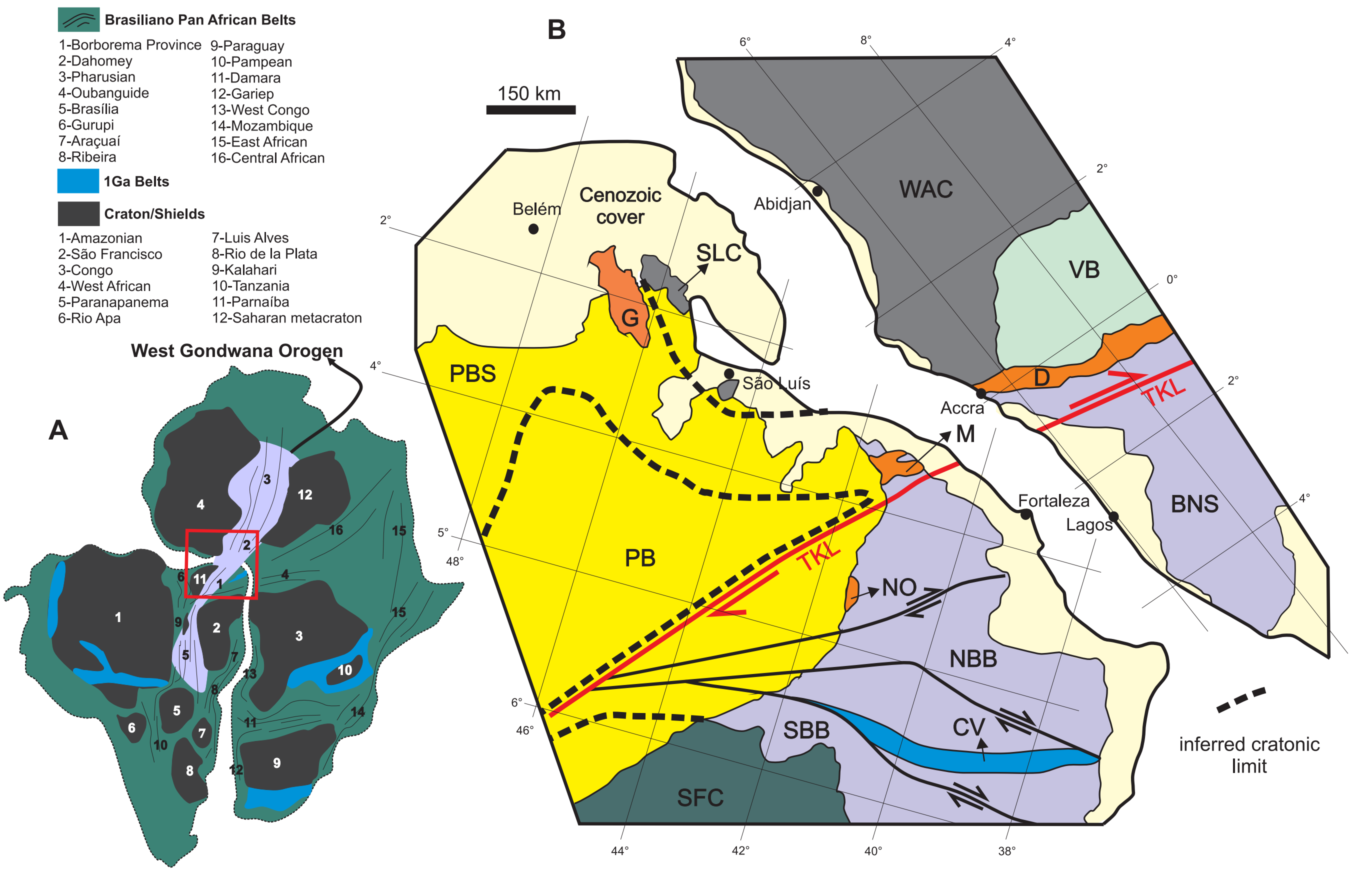

Fig. 1. Continental context of the investigated area. (A) Extent of the West Gondwana Orogen and position of the main cratonic blocks. (B) Equatorial Brazil–Africa correlation modified from Klein and Moura (2008). WAC: West African craton; VB: Volta basin; D: Dahomey belt; BNS: Benino-Nigerian Shield; SLC: São Luiz craton; G: Gurupi belt; M: Martinópole Group; NO: Novo Oriente basin; NBB: Northern Borborema block; SBB: Southern Borborema block; SFC: São Francisco craton; CV: Cariris Velhos belt; PB: Parnaíba block; PBS: Parnaíba basin, TKL: Transbrasiliano-Kandi Lineament.

Modern and traditional views suggest that the WGO resulted from the consumption and closure of the Goiás-Pharusian Ocean that culminated in a continent–continent collision involving mainly the conjoined Amazonian and West African cratons plus the concealed Parnaíba Block colliding against the São Francisco and Saharan cratons (e.g. Trompette, 1994; Cordani et al., 2003; Ganade de Araujo et al., 2014a,b).

With a protracted tectonic history of over 400 m.y. the WGO records accretionary convergent tectonics since the Early Neoproterozoic with development of several intra-oceanic and continental arcs that are now preserved within a deeply eroded, paleo-collisional zone (e.g. Caby, 1989; Berger et al., 2011; Ganade de Araujo et al., 2014a,c). The dominant rock types caught up between the cratonic blocks are interpreted as passive margin deposits, early juvenile and late evolved continental-arc rock assemblages and syn-orogenic supracrustal sequences. Moreover, the orogen has all key features that are observed in modern collision zones with continental subduction occurring nearly simultaneously over at least a 2500-km-long section of the orogen during the Ediacaran period (610–620 Ma) (Ganade de Araujo et al., 2014a).

Here we review the tectonic evolution of the Dahomey belt of the WGO and discuss new geochronological and isotopic zircon data in light of similar data previously acquired for the geologically related Northern Borborema Province, in NE Brazil. This area records a complete cycle of plate tectonics and is not only important because of its crucial position for Gondwana paleogeographic recontructions, but also because it is an open laboratory exposing the deep roots of Himalayan-like mountains.

2. Geological outline of the Dahomey belt and Volta basin

The Dahomey belt resulted from collision between the passive continental margin of the West Africa Craton (WAC) and the eastern continental block known as the Benino-Nigerian Shield (Caby, 1987; Castaing et al., 1994; Attoh et al., 1997; Bessoles and Trompette, 1980; Affaton et al., 1991) (Figs. 1 and 2). The connection of this shield to the proposed larger Saharan metacraton (Abdelsalam et al., 2002) further east is tantalizing, but is yet to be demonstrated. The belt corresponds to the southern segment of the Pan-African Trans-Saharan orogeny that extends for >2500 km from the Sahara desert to the Gulf of Guinea (Caby, 1987). Prior to the opening of the Atlantic Ocean this large belt was connected to the orogenic areas of northeast and central Brazil running for more than 4000 km in the Neoproterozoic West Gondwana Orogen (Caby, 1989; Trompette, 1994; Cordani et al., 2013a,b; Ganade de Araujo et al., 2014a).

In Ghana and adjoining parts of Togo and Benin, the Dahomey belt has a well-organized orogenic architecture with passive margin-related rocks, belonging to the WAC dominating in the external (westerly) portion of the orogen and active margin-related rocks dominating its internal (easterly) portion, marking the western active margin of the Benino-Nigerian shield (Affaton et al., 1991; Agbossoumondé et al., 2004; Attoh and Nude, 2008).

The classical subdivision of the Dahomey belt (Affaton, 1990) includes three main structural units: (i) the western external structural units corresponding to the Buem and Atacora groups; (ii) the eastern internal structural units (Benino-Nigerian shield basement) and, (iii) in between the two, the so-called Dahomey

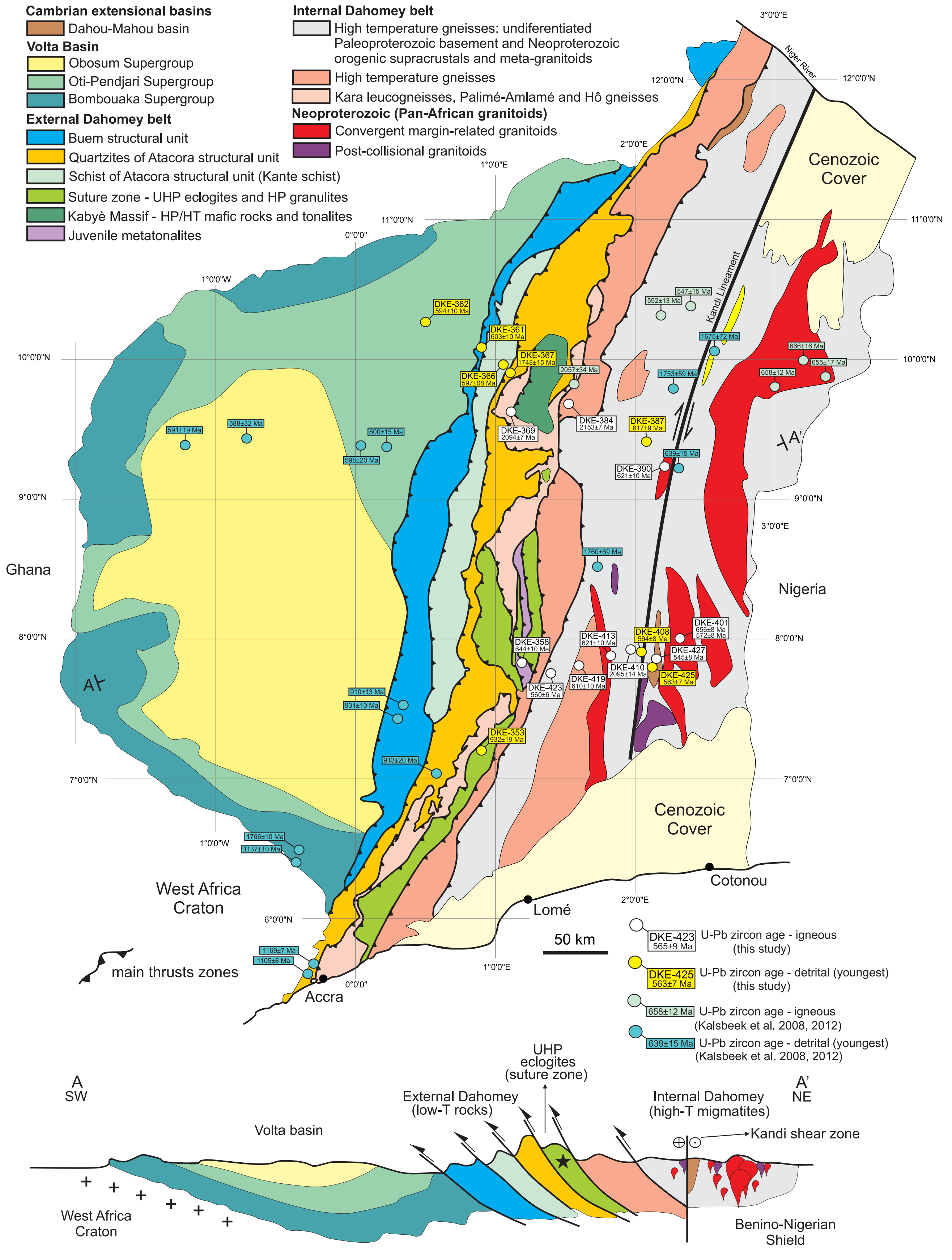

Fig. 2. Geological map of Ghana, Togo and Benin illustrating the context of the Volta Basin, the Dahomey belt and the Benino-Nigerian Shield. Schematic cross-section illustrating the west-verging thrusts of the Dahomey belt upon the flat-lying sedimentary rocks of Volta Basin.

Source: Modified from Sylvain et al. (1986).

suture zone is characterized by high-grade mafic–ultramafic massifs with some remnants of UHP eclogites. The Buem and Atacora structural units correspond to tectonic collages of various lithologies including both metasedimentary rocks from the WAC passive margin and pre-Neoproterozoic gneisses representing the old basement.

The easterly Buem Structural Unit has been thrusted to the west onto the Volta Basin and its basement (WAC) (Fig. 2). It consists of sedimentary or weakly metamorphosed sedimentary rocks (shales, siltstones, quartzitic sandstones, hematitic rocks and sericite schists) associated with mafic volcanics and serpentinized ultramafic rocks (Affaton et al., 1997). The Atacora Structural Unit represents a thick pile of nappes composed of schists and quartzites that are overthrusted onto the Buem Structural Unit (Affaton, 1990) (Fig. 2). It tectonically underlies the external nappes of the reworked basement complex (the Kara-Niamtougou or Mô orthogneissic units, the Sokodé-Kéméni Unit, the plutono-metamorphic Palimé-Amlamé and/or Hô Unit) (Affaton, 1990; Agbossoumondé et al., 2007).

The suture zone is well exposed from southeast Ghana to northwest Benin and corresponds to a narrow and lithologically diverse area marked by striking positive gravity and magnetic anomalies (El-Hadj Tidjani et al., 1997). Numerous ultramafic and mafic units are scattered throughout the suture zone and display contrasting lithological and metamorphic features hosting UHP and HP eclogites, HP granulites and amphibolites (Attoh et al., 1997; Agbossoumondé et al., 2001, 2004; Duclaux et al., 2006; Ganade de Araujo et al., 2014a). Along this suture zone, deformed alkaline rocks, carbonatite and ultrahigh-pressure (UHP) rocks occur (Attoh and Nude, 2008; Nude et al., 2009; Ganade de Araujo et al., 2014a) in association with high-pressure granulites and eclogites of basaltic composition (Attoh, 1998; Attoh and Morgan, 2004; Agbossoumondé et al., 2013). In some places, the UHP rocks are interleaved with quartzites of the Atacora Structural Unit, hence indicating subduction of sections of the WAC passive margin to mantle depths at ca. 610 Ma (the age of the UHP event, Ganade de Araujo et al., 2014a). In this convergent collisional context, the structural arrangement, with west-verging thrusts of the western external units, suggests that the WAC was the underthrusted continental plate and that the eastern internal basement rocks mark the western margin of the overriding Benino-Nigerian Shield.

The eastern internal basement rocks are made of granitoid gneiss, migmatite assemblages comprising the Accra and Benin Plains units that account for much of the Benino-Nigerian Shield (Attoh et al., 2013; Aidoo et al., 2014). The Benino-Nigerian Shield represents an association of high-temperature granulites, amphibolite facies gneisses, metasedimentary rocks, migmatites and granitoid gneisses, which are thought to be part of the WAC (Agbossoumondé et al., 2007). These rocks are cross-cut by various types of granitoids and also host younger low-grade metavolcanic and sedimentary formations (e.g. Daho-Mahou Basin, Fig. 2).

The Dahomey belt together with the Benino-Nigerian Shield are correlated with the Brasiliano (ca. 600 Ma) Borborema Province of NE Brazil. Both regions are essentially Palaeoproterozoic terranes, strongly reworked by Neoproterozoic tectonometamorphic and magmatic activity (Abdelsalam et al., 2002; Caby, 1989; Arthaud et al., 2008; Santos et al., 2008; Dada, 2008; De Wit et al., 2008b, and references therein). Most of these correlations assume that the Transbrasiliano Lineament in South America continues into the Trans-Saharan of West Africa as the Kandi Lineament. The dextral Transbrasiliano-Kandi Lineament forms a long (up to 4000 km) strike-slip belt that resulted from relative movement obliquity during continental collision. This strike-slip belt reactivated previous sutures, and dextral movement started possibly as early as 610 Ma, soon after onset of continental collision (Ganade de Araujo et al., 2014a,b).

The Kandi Lineament represents a wide (up to 50 km width), N-S trending dextral transcurrent shear zone, where the deformation occurs under high to low-temperature conditions (Adissin, 2012). Recently, Adissin (2012) obtained U–Pb zircon ages of 2091 ± 14 Ma and 2057 ± 8 Ma from metamorphic rocks within the Kandi Lineament (granulite and amphibolite facies gneisses) and interpreted these ages as the crystallization and metamorphic imprint at granulite facies conditions, respectively. The amphibolite facies gneiss yielded a Neoproterozoic age of 606 ± 5 Ma that corresponds to the mylonitic deformation associated to the shearing at those metamorphic conditions.

The sedimentary sequence of the Volta Basin lies unconformably on the southeastern margin of the WAC. This basin is truncated to the east by the thrust front of the Dahomey belt (Affaton, 2008). The basin is divided into three unconformity-bounded sequences. The lower sequence or Bombouaka Supergroup comprises a 500 m thick package of sandstones and siltstones accumulated in deltaic, fluvial and nearshore or shoreface environments (Carney et al., 2010). The middle sequence or Oti-Pendjari Supergroup comprises a succession that records the transition from shallow marine environments, in the Kodjari Formation, to a marine foreland basin sequence represented by argillaceous strata interbedded with highly immature, wacke-type sandstones and conglomerates of the Afram and Bimbila formations (Carney et al., 2010). This unit thickens eastwards to an estimated 3–4 km, and in the same direction becomes increasingly flexured towards the frontal thrust systems of the Dahomey belt (Bertrand-Sarfati et al., 1991; Affaton, 2008; Carney et al., 2010). The upper megasequence or Obosum Group (or Tamale Supergroup), consists of distinctive lithic, feldspar-rich arenites and conglomerates deposited as terrestrial molasse during the final stages of uplift within the Dahomey belt (e.g. Bertrand-Sarfati et al., 1991; Kalsbeek et al., 2008; Carney et al., 2010).

3. Sampling and analytical procedures

In order to investigate timing of orogenic processes in the Dahomey belt we have collected 18 samples from different tectonic settings and regions within the orogen for coupled U–Pb/Lu–Hf zircon isotopic investigations. A total of 11 igneous and metaigneous rocks and 7 sedimentary and/or metasedimentary rocks distributed orthogonally to the belt's trend were collected (Fig. 2). Sample site locations and a brief summary of the samples are provided in Table 1.

Zircon grains were separated from fresh crushed rocks (3–5 kg) using conventional and heavy liquid and magnetic techniques (jaw crusher, disk grinder, Wilfley table, Frantz isodynamic magnetic separator and density separation using bromoform and methylene iodite). Around 50–80 zircon grains from each sample were mounted in epoxy resin, polished to half of mean grain thickness for further imaging with transmitted light and cathodoluminescence to unravel internal complexities. Cathodoluminescence (CL) images of zircon grains were obtained using a Quanta 250 FEG electron microscope equipped with Mono CL3 + cathodoluminescence spectroscope (Centaurus) at the Geochronological Research Center in São Paulo University, Brazil.

U–Pb analyses were done using SHRIMP IIe at the Geochronological Research Center (CPGeo) at the São Paulo University. The data have been reduced in a manner similar to that described by Williams (1998 and references therein), using the SQUID Excel Macro of Ludwig (2001). Uncertainties given for individual U–Pb analyses (ratios and ages) are at the 1σ level, however uncertainties in the calculated weighted mean ages are reported as 2σ confidence limits. For the age calculations, corrections for common Pb were made using the measured ^{204}Pb and the relevant common Pb compositions from the Stacey and Kramer (1975) model.

Table 1
Site location and summary of the data of the investigated samples.

Sample	Coordinates (UTM, zone 31)	Lithology	Unit	Age (Ma)
DKE-369	292995;1064476	Muscovite-rich leucocratic gneiss	Kara gneiss	2094 ± 6.8
DKE-384	330475;1073542	Granodioritic schollen	Benino-Nigerian Shield	2153 ± 6.5
DKE-410	386412;876029	Granulitic mylonitic granodioritic orthogneiss	Benino-Nigerian Shield	2095 ± 14
DKE-401A	421788;883373	Patchy granodioritic migmatite	Benino-Nigerian Shield	655.7 ± 8.2
DKE-401B	421788;883373	Felsic dike	Benino-Nigerian Shield	572.4 ± 8.1
DKE-358	300500;865897	Hornblende-biotite mafic meta-tonalite	Benino-Nigerian Shield	643.7 ± 9.8
DKE-390	411800;1021028	Hornblende granodioritic orthogneiss	Benino-Nigerian Shield	621.6 ± 9.5
DKE-413	370586;870509	Biotite-hornblende orthogneiss	Benino-Nigerian Shield	621 ± 9.4
DKE-419	342704;862916	Biotite-hornblende orthogneiss	Benino-Nigerian Shield	610 ± 9.4
DKE-423	323325;857419	Migmatized biotite granodioritic orthogneiss	Benino-Nigerian Shield	560 ± 5.8**
DKE-427	401875;871810	Fine-grained dacite	Daho-Mahou basin	545.7 ± 7.8
DKE-362	225296;1136976	Reddish beige argillite	Oti-Pedjari Group	594 ± 10*
DKE-361	268423;1117505	Coarse-grained immature sandstone	Buem Group	603 ± 10*
DKE-366	287032;1101824	Chlorite-sericite schist	Kante Unit	597 ± 8*
DKE-367	290696;1096370	Muscovite quartzite	Atacora Group	1748 ± 15*
DKE-353B	258071;792210	Phengite-rich quartzite	Atacora Group	932 ± 19*
DKE-387	399792;1040334	Metasedimentary schollen	Benino-Nigerian Shield	617 ± 09*
DKE-425	402891;861471	Polymictic matrix-supported conglomerate	Daho-Mahou	563 ± 7*

* Youngest detrital zircon.
** Age of anatexis.

Concordia plots, regressions and any weighted mean age calculations were carried out using Isoplot/Ex 3.0 (Ludwig, 2003) and where relevant include the error in the standard calibration. U–Pb geochronological results are presented in Table S1 of Supplementary data.

For the detrital zircon grains U–Pb analyses were acquired by a Neptune laser-ablation multi-collector inductively coupled plasma mass spectrometer (LAM-ICP-MS) coupled to an excimer ArF laser (λ = 193 nm) ablation system also at the Geochronology Research Center of the São Paulo University. The mounts containing zircon grains were cleaned in a HNO_3 solution (3%) and in ultraclean water bath. The ablation was done with spot size of 29 μm, at a frequency of 6 Hz and an intensity of 6 μJ. The ablated material was carried by Ar (~0.7 L/min) and He (~0.6 L/min) in analyses of 60 cycles of 1 s. Unknowns were bracketed by measurements of the international standard GJ-1, following the sequence 2 blanks, 3 standards, 13 unknowns, 2 blanks, and 2 standards. Raw data were reduced using a home-made spreadsheet and corrections were done for back-ground, instrumental mass bias drift and common Pb. The ages were calculated using ISOPLOT 3.0 (Ludwig, 2003).

Lu–Hf analyses were also carried out at the Geochronological Research Center (CPGeo) at the São Paulo University also using the same described LAM-ICP-MS equipped with a laser system. The laser spot used was 39 μm in diameter with an ablation time of 60 s, repetition rate of 7 Hz, and He used as the carrier gas (Sato et al., 2009). $^{176}Hf/^{177}Hf$ ratios were normalized to $^{179}Hf/^{177}Hf = 0.7325$. The isotopes ^{172}Yb, ^{173}Yb, ^{175}Lu, ^{177}Hf, ^{178}Hf, ^{179}Hf, ^{180}Hf, and $^{176}(Hf + Yb + Lu)$ were simultaneously measured. $^{176}Lu/^{175}Lu$ ratio of 0.02669 was used to calculate $^{176}Lu/^{177}Hf$. Mass bias corrections of Lu–Hf isotopic ratios were done applying the 176-11 variations of GJ1 standard. A decay constant for Lu of 1.867 × 10 (Söderlund et al., 2004), the present-day chondritic ratios of $^{176}Hf/^{177}Hf = 0.282772$ and $^{176}Lu/^{177}Hf = 0.0332$ (Blichert-Toft and Albarède, 1997) were adopted to calculate εHf values. A two-stage continental model (TDM) was calculated using the initial $^{176}Hf/^{177}Hf$ of zircon and the $^{176}Lu/^{177}Hf = 0.022$ ratio for the lower continental crust (Griffin et al., 2004). Zircon Lu–Hf isotopic results are presented in Table S2 of Supplementary data.

4. Results

Isotopic results for samples of the Dahomey belt from Togo and Benin are available in the supplementary data (Tables S1 and S2). Crystallization ages for igneous and metaigneous rocks were acquired by SHRIMP technique whereas detrital zircon grains from sedimentary and metasedimentary rocks were dated by LAM-ICP-MS. Lu–Hf isotopic measurements via LAM-ICP-MS were all carried out on the same textural domain in each zircon previously analysed by the U–Pb SHRIMP technique.

4.1. *Zircon U–Pb ages and Hf isotopes from igneous and metaigneous rocks*

4.1.1. *Metaigneous rocks from the basement of the Benino-Nigerian Shield*

Three metaigneous rocks from the deformed gneissic basement were collected in the internal zone of the orogen in the Benino-Nigerian Shield.

Sample DKE-410 is a granulitic mylonitic granodioritic orthogneiss collected in a quarry in the vicinity of the Kandi Lineament close to the city of Savalou. Zircon grains were extracted from a strongly foliated portion, avoiding contamination with any leucocratic vein. They range in size from 80 to 250 μm and have length to width ratios from 2:1 to 4:1. Cathodoluminescence images reveal a well-developed oscillatory zoning typical of magmatic zircon crystals, surrounding a low-U core (c.f. zircon 6.1 – Fig. 4). Thirteen analyses were done on the zircon crystals, which yielded a calculated upper intercept age of 2095 ± 14 Ma. Three most concordant analyses yielded a mean age of 2086 ± 7 Ma (2σ), interpreted as the crystallization age of the granodioritic protolith (Fig. 5A).

Sample DKE-369 consists of a muscovite-rich leucocratic gneiss (Kara orthogneiss) collected close to the locality of Tchitchao. Zircons range in size from 80 to 300 μm and have length to width ratios ranging from 2:1 to 5:1. Cathodoluminescence images reveal a well-developed oscillatory zoning typical of magmatic zircons (c.f. zircon 5.1 – Fig. 4). Analyses on selected zircons were invariably discordant and defined a discordia line (n = 17) that yielded an upper intercept age of 2094 ± 6.8 Ma (2σ), interpreted as the best estimate of the crystallization age of the gneiss protolith, essentially within error of the previous sample (Fig. 5B).

Sample DKE-384 is a granodioritic schollen embedded in diatexite migmatite collected close to Sonoumon-Barierow. Zircon crystals range in size from 80 to 250 μm and have length to width ratios from 2:1 to 4:1. Cathodoluminescence images reveal a well-developed oscillatory zoning typical of magmatic zircon (Fig. 4), however some grains display structures typical of magmatic resorbed zircon. Twenty-five analyses yielded an upper intercept age of 2146 ± 17 Ma (2σ) (Fig. 5C). A cluster of analyses near

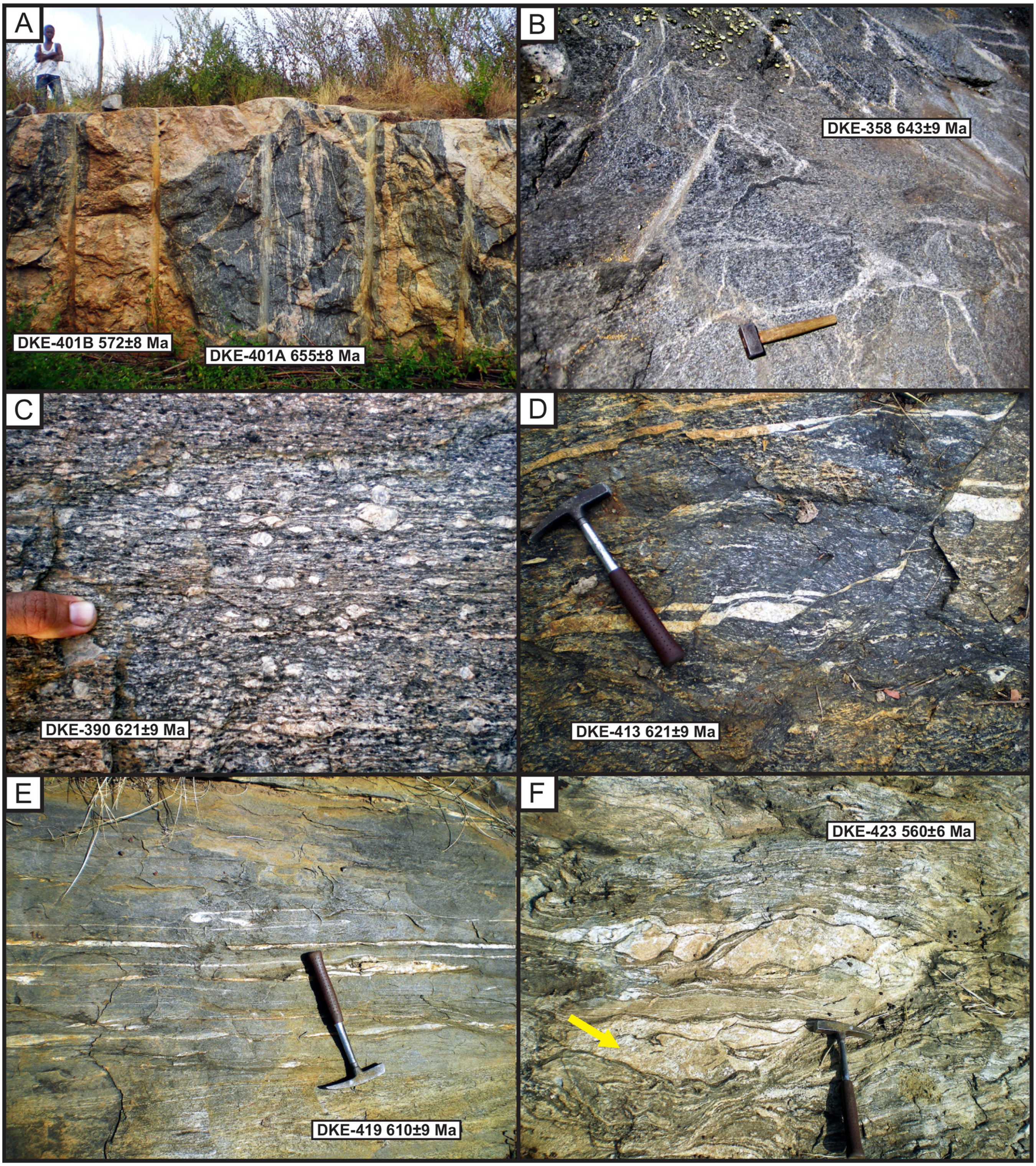

Fig. 3. Field photographs of the studied rocks from the Benino-Nigerian Shield. (A) Patchy granodioritic migmatite (sample DKE-401A) cut by felsic quartz–feldspathic dyke (sample DKE-401B) interpreted as *in source* leucosomes. (B) Partially melted juvenile hornblende-biotite mafic meta-tonalite (sample DKE-358). (C) Mylonitic Hornblende granodiorite orthogneiss (sample DKE-390) close to the locality of Kika. (D) Mylonitic biotite-hornblende orthogneiss locally migmatized in the Zou River (Sample DKE-413). (E) Biotite-hornblende orthogneiss locally migmatized in the Okpe River (Sample DKE-419). (F) Migmatized biotite granodioritic orthogneiss (sample DKE-423, yellow arrow).

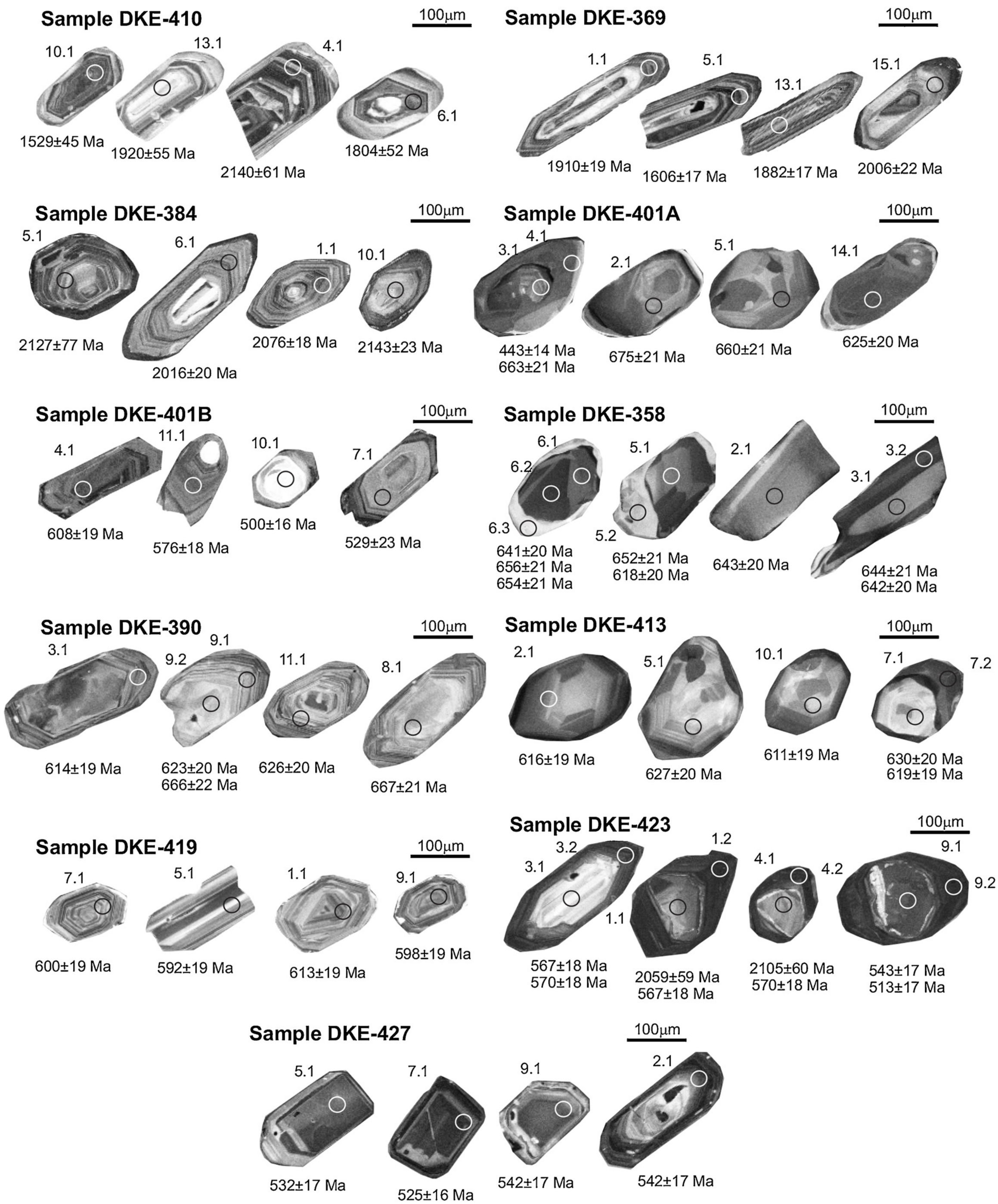

Fig. 4. Selected zircon crystals from igneous rocks investigated through SHRIMP and LAM-ICP-MS.

the upper intercept yielded a weighted mean concordia age of 2153 ± 6.5 Ma (2σ), interpreted as the best estimate of the crystallization age of the granodioritic protolith (Fig. 5D).

4.1.2. Neoproterozoic magmatism in the Benino-Nigerian Shield

The Benino-Nigerian Shield in Togo and Benin is intruded by several granitic (s.l.) bodies of different compositions, shapes and sizes. Due to the lack of systematic mapping, many of these bodies are not yet properly individualized and they mix up with the Paleoproterozoic igneous rocks and younger Neoproterozoic supracrustal rocks. We collected seven different types of granitoids (deformed and undeformed) including anatectic granitoids, attempting to bracket timing of Neoproterozoic igneous activity and crustal reworking in the Benino-Nigerian Shield.

Samples from site DKE-401, in the vicinities of Thio (Fig. 2), are divided into a patchy granodioritic migmatite (sample DKE-401A) and a crosscutting felsic quartz–feldspathic dykelet (sample DKE-401B) (Fig. 3A). These dykelets could be related to the anatectic event seen in the granodiorite in the form of narrow connected leucocratic veins and pockets of melt or could consist of transported melt from elsewhere. Zircon crystals from the granodioritic paleosome are euhedral, transparent, and colourless to light yellow. Most of them are equant to short prismatic. Crystals range in length from 80 to 250 μm. Most zircon crystals are oscillatory or sector zoned and interpreted as the result of magmatic growth

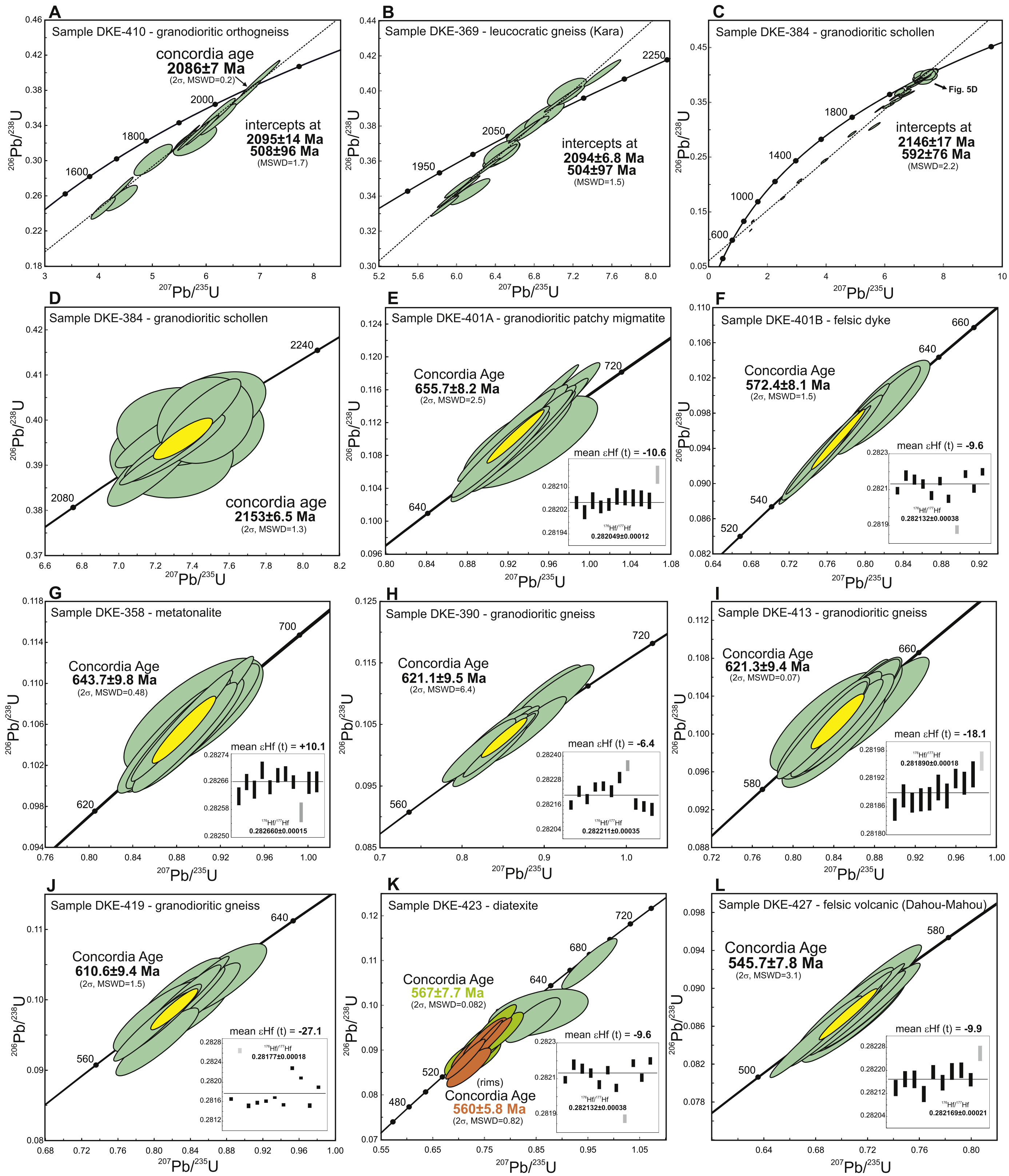

Fig. 5. Zircon U–Pb SHRIMP geochronology and Lu–Hf isotopes (insets) of the igneous rocks of the Dahomey belt and Benino-Nigerian Shield.

(c.f. zircon 5.1 – Fig. 4), however some zircon crystals have a nicely preserved core (c.f. zircon 3.1 – Fig. 4). Fourteen zircon crystals with Th/U ratios of 0.57–1.04 form a group in the concordia line yielding a mean age of 655.7 ± 8.2 Ma (2σ) for the granodioritic protolith crystallization (Fig. 5E). Calculated (^{176}Hf/^{177}Hf)$_t$ ratios from these zircon crystals have a narrow variation between 0.282010 and 0.282138 which corresponds to εHf$_{(t)}$ values between −12.0 and −7.6. Zircon crystals from the felsic quartz–feldspathic vein (sample DKE-401B) are also euhedral to subhedral and have well-defined igneous oscillatory zoning with Th/U ratios from 0.06 to

0.57. Some zircon crystals have a core-rim structure with cores being U-poorer than the rims (c.f. zircons 10.1 and 11.1 – Fig. 4). A concordia age of 572.4 ± 8.1 Ma (2σ) defined by 14 concordant analyses reflects the age of crystallization of this granitic vein (Fig. 5F). Calculated $(^{176}Hf/^{177}Hf)_t$ ratios from zircon crystals from this sample (DKE-401B) vary between 0.281872 and 0.282192, which corresponds to $\varepsilon Hf_{(t)}$ values between −13.2 and −8.2.

Sample DKE-358 is a mafic, hornblende-biotite meta-tonalite (Fig. 3B) collected along the road to Rodokpe in the limit between the internal Benino-Nigerian Shield and the external units of the Dahomey belt (Fig. 2). In general, zircon crystals from this sample range in size from 80 to 250 μm and have length to width ratios ranging from 2:1 to 4:1. Cathodoluminescence images reveal a well-developed sector to oscillatory zoning typical of magmatic zircons, but some display a comparative low-U rim (c.f. zircons #5.2 and #6.3 – Fig. 4). Analysed zircons crystals have U contents between 60 and 341 ppm and Th/U ratios ranging mainly from 0.58 to 0.99. Twelve analyses yielded a concordia age of 643.7 ± 9.8 Ma (2σ), interpreted as the crystallization age of the tonalitic protolith (Fig. 5G). In contrast to the previous sample, zircon crystals have a narrow variation of $(^{176}Hf/^{177}Hf)_t$ with values ranging from 0.282613 to 0.282568, which corresponds to $\varepsilon Hf_{(t)}$ values between +7.2 and +10.8 (Fig. 7).

Sample DKE-390 is a mylonitic hornblende granodioritic orthogneiss (Fig. 3C) with K-feldspar porphyroclasts displaying a right-lateral shear sense and collected close to the Kandi Lineament in the locality of Kika in Benin (Fig. 2). Zircon grains were extracted from the granodioritic paleosome, avoiding contamination with the neosome. In general they are subhedral to euhedral, translucent and colourless, with dimensions ranging from 60 to 250 μm. They have a complex and well-developed oscillatory zoning characteristic of igneous zircon (c.f. zircon #11.1 – Fig. 4). Th/U ratios range from 0.51 to 0.83. A concordia age defined by 10 concordant analyses yielded an age of 621.6 ± 9.5 Ma (2σ), interpreted as the crystallization age of the granodioritic protolith (Fig. 5H). Calculated $(^{176}Hf/^{177}Hf)_t$ ratios from the analysed zircons vary from 0.282138 to 0.282350 with $\varepsilon Hf_{(t)}$ varying from −8.7 to −3.4.

Sample DKE-413 consists of a biotite-hornblende mylonitic orthogneiss, from the Benino-Nigerian Shield, collected in the Zou river. It has a more mafic composition in comparison to the previous sample (Fig. 3D), and is also affected by the right-lateral shear of the Kandi Lineament. Zircon crystals from the protolith are euhedral to subhedral with ovoid shapes and sizes ranging from 50 to 150 μm. Most zircon crystals are sector and/or oscillatory zoned with some displaying magmatic resorbed overgrowths (c.f. zircons #7.1 and #7.2 – Fig. 4). In general Th/U ratios vary from 1.37 to 0.75. A concordia age of 621 ± 9.4 Ma (2σ) was defined by 13 concordant analyses and reflects the age of crystallization of the protolith (Fig. 5I). Calculated $(^{176}Hf/^{177}Hf)_t$ ratios from the analysed zircon crystals vary from 0.281846 to 0.281954 with $\varepsilon Hf_{(t)}$ varying from −15.5 to −19.2.

Sample DKE-419 collected close to the Okpe river (Fig. 3E) also has a similar composition to the previous two samples, consisting of a biotite-hornblende orthogneiss locally migmatized. In general, zircon crystals have a well-developed igneous oscillatory zoning surrounded by a metamorphic overgrowth too thin to be analysed (c.f. zircon 9.1 – Fig. 4). The dated igneous zircons have Th/U ratios of 0.43–2.81 and yielded a 12-point concordia age of 610 ± 9.4 Ma (2σ) that reflects the age of crystallization of the tonalite (Fig. 5J). Calculated $(^{176}Hf/^{177}Hf)_t$ ratios from these zircons have a significant variation between 0.281489 and 0.281895 which corresponds to $\varepsilon Hf_{(t)}$ values between −32.1 and −17.4.

Sample DKE-423, collected close to the Ouanoukope village (Fig. 3E), is a strongly migmatized biotite granodioritic orthogneiss with characteristic diatexitic sectors. Zircon crystals were extracted from the diatexite portion (Fig. 3E) and are euhedral to subhedral with sizes ranging from 50 to 250 μm. Most zircon crystals have a well-developed rim surrounding inherited cores (c.f. zircons #3.1 and #3.2 – Fig. 4). Th/U ratios do not vary systematically according to zircon structure, with most values ranging from 0.10 to 0.84. Two Paleoproterozoic inherited zircon grains yielded $^{207}Pb/^{206}Pb$ ages of 2127 ± 14 and 2080 ± 7 Ma. Among the Neoproterozoic zircon crystals identified as cores, a cluster defines a concordia age of 567 ± 7.7 Ma (2σ) whereas rims surrounding these cores yielded a concordia age at 560 ± 5.8 Ma (2σ) (Fig. 5K). Older Neoproterozoic zircon crystals with $^{206}Pb/^{238}U$ ages of 669 ± 21, 605 ± 20 and 595 ± 19 Ma probably represent an inherited component. We interpret the age of 560 ± 5.8 Ma derived from the rims as the best estimate for the timing of the latest partial melting event in this rock.

4.1.3. Cambrian volcanism in the extensional Dahou-Mahou basin

A number of extensional troughs developed over the older basement rocks and the Neoproterozoic crystalline rocks in the Benino-Nigerian Shield; the Dahou-Mahou being the most prominent. Sample DKE-427 collected in this basin close to the locality of Savalou consists of a fine-grained dacite with K-feldspar phenocrysts. Zircon crystals from this volcanic rock are mostly euhedral (80–150 μm) and characterized by a prominent oscillatory zoning (c.f. zircons 2.1 and 5.1 – Fig. 4). Th/U ratios for the dated zircon crystals vary significantly from 0.43 to 1.29. Ten concordant analyses fall in a group yielding a concordia age of 545.7 ± 7.8 Ma (2σ) that reveals the age of the volcanism in the basin (Fig. 5L). Calculated $(^{176}Hf/^{177}Hf)_t$ ratios from the analysed zircon crystals range from 0.282109 to 0.282248 that correspond to $\varepsilon Hf_{(t)}$ of −11.6 to −6.5.

4.2. Zircon U–Pb ages from (meta)-sedimentary rocks of the Dahomey belt

We performed 466 analyses on detrital zircon grains from metasedimentary rocks of the internal Benino-Nigerian Shield, external units representing passive margin sediments of the WAC (Atacora and Buem Groups) of the Dahomey belt in addition to the sedimentary rocks of the Volta and Dahou-Mahou basin. Among them 344 yielded reliable ages with less than 5% discordance ($^{206}Pb/^{238}U$ to $^{207}Pb/^{235}U$) and 10% discordance ($^{206}Pb/^{238}U$ to $^{207}Pb/^{206}Pb$).

4.2.1. High-grade supracrustal rocks within the Benino-Nigerian Shield

With rare preserved quartzitic crests, supracrustal rocks of the Benino-Nigerian Shield are dominated by high-temperature migmatites. Here we investigated two samples representative of these high-grade gneisses.

Sample DKE-387 consists of a metasedimentary schollen embedded in diatexite migmatite containing abundant residual sillimanite and garnet, collected in the Benino-Nigerian Shield area (Fig. 2). Zircon grains were extracted from the schollen avoiding direct contamination with the neosome. They are euhedral, transparent, and colourless to light yellow. Most of them are equant to short prismatic. Crystals range in length from 80 to 200 μm. Most of the detrital zircon grains are oscillatory zoned and interpreted as the result of magmatic growth in the source rock, but despite of the effort to avoid contamination with the neosome, the newly developed rims around magmatic cores also with a characteristic oscillatory zoning are interpreted as melt-precipitated zircon. The youngest concordant analysed detrital zircon grain yielded a $^{206}Pb/^{238}U$ age of 617 ± 09 Ma with the youngest population clustering at ca. 605 Ma. Neoproterozoic grains (47%) range mainly from 596 to 781 Ma with a pronounced peak at ca. 690 Ma. One

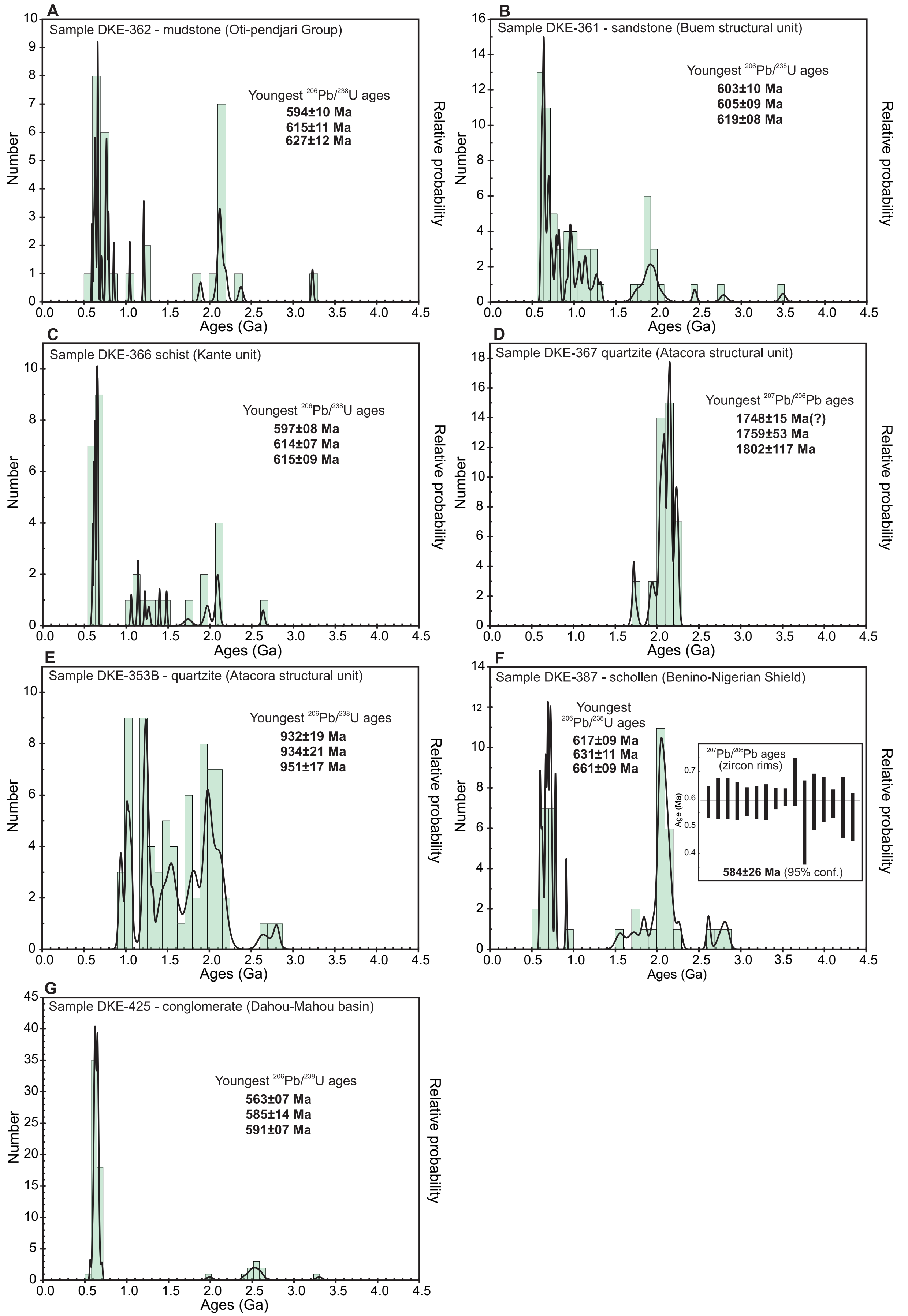

Fig. 6. Detrital zircon U–Pb LAM-ICP-MS geochronology from the metasedimentary and sedimentary rocks of the Dahomey belt and Benino-Nigerian Shield.

Mesoproterozoic zircon grain had a $^{207}Pb/^{206}Pb$ age of 1555 ± 5. Paleoproterozoic zircons (51%) range from 1720 to 2268 with a well-defined peak at 2050 Ma. Three Archaean zircon grains were analysed and the oldest one yielded a concordant $^{207}Pb/^{206}Pb$ age of 2834 ± 38 Ma (Fig. 6F). The melt-precipitated overgrowths (n = 16) yielded a $^{207}Pb/^{206}Pb$ weighted mean age of 584 ± 26 Ma (95% conf.), which is interpreted as the age of the partial melting event in this sample (inset in Fig. 6F).

4.2.2. Metasedimentary rocks of the external Dahomey belt (passive margin of the WAC)

Contrary to the internal Benino-Nigerian Shield, supracrustal rocks of the external zone of the Dahomey belt are dominated by low-temperature to locally high-pressure siliciclastic rocks, with abundant quartzitic packages of the Atacora Unit (Fig. 2). The metamorphic grade diminishes towards the west where sandstones of the Buem Unit dominate.

Sample DKE-361 from the Buem structural unit comprises a greenish beige, medium to coarse-grained, immature sandstone, collected in the Oumbé River. From the 90 analysed zircon grains 63 yielded reliable ages. The youngest zircon yielded a $^{206}Pb/^{238}U$ age of 603 ± 10 Ma with the youngest population clustering at ca. 605 Ma. Of the total dated zircon grains, 60% show Neoproterozoic ages spanning from 980 to 603 Ma, but with a well-defined peak at ca. 634 Ma. Mesoproterozoic ages range from 1030 to 1319 Ma (15%) while Paleoproterozoic zircon grains vary from 2304 to 1710 Ma (19%) with a main peak at ca. 1876 Ma. Two Archaean zircon grains were analysed and the oldest one yielded a concordant $^{207}Pb/^{206}Pb$ age at 3497 ± 34 Ma (Fig. 6B).

Sample DKE-366 from the Atacora structural unit collected close to Kante (Kante schist) consists of a green chlorite-sericite schist. From the 71 analysed zircon grains only 30 yielded useful ages. The youngest zircon yielded a $^{206}Pb/^{238}U$ age of 597 ± 8 Ma with the youngest population clustering at ca. 610 Ma. Of the total dated zircon grains, 50% show Neoproterozoic ages spanning from 662 to 597 Ma with a well-defined peak at ca. 649 Ma. Mesoproterozoic (23%) and Paleoproterozoic zircons (23%) range from 1482 to 1059 Ma and from 2038 to 1704 Ma, respectively. One Archaean zircon grain was analysed yielding a concordant $^{207}Pb/^{206}Pb$ age of 2635 ± 38 Ma (Fig. 6C).

Sample DKE-367 from the Atacora structural unit consists of a muscovite quartzite collected in the main road to Kara. Neoproterozoic zircon grains were not found in this sample, which is dominated only by Paleoproterozoic zircon grains with the youngest yielding a $^{207}Pb/^{206}Pb$ age of 1759 ± 53 Ma (Fig. 6D).

Sample DKE-353B consists of a phengite-rich quartzite related to the UHP eclogites of the Lato Hills and is also related to the Atacora structural unit. From the 77 analysed zircons 72 yielded reliable ages. The youngest concordant analysed zircon grain yielded a $^{206}Pb/^{238}U$ age of 932 ± 19 Ma with the youngest population clustering at ca. 940 Ma. Only four grains have Neoproterozoic ages ranging from 993 to 932 Ma. Mesoproterozoic zircon grains (42%) are abundant spanning from 1006 to 1598 Ma. Paleoproterozoic zircon grains (40%) are also rather abundant comprising ages from 2225 to 1730 Ma with a well-defined peak at ca. 1965 Ma. Three Archaean zircon grains were analysed and the oldest one yielded a concordant $^{207}Pb/^{206}Pb$ age of 2805 ± 32 Ma (Fig. 6E).

4.2.3. Sedimentary rocks of the Volta and Dahou-Mahou basins

The Volta foreland basin and the Dahou-Mahou post-collisional basin consist of non- to weakly metamorphosed siliciclastic rocks deposited after the main orogenic peak of the Dahomey belt. Kalsbeek et al. (2008) reported abundant detrital zircon ages from the sedimentary rocks of the Volta basin and thus only one sample from this basin was investigated in this study.

Sample DKE-362 from the Oti-Pedjari Supergroup of the Volta basin is a reddish beige mudstone collected along the road to Kpesside. Forty-seven zircon grains from this sample were analysed, but only 30 provided reliable ages. The youngest concordant analysed zircon grain yielded a $^{206}Pb/^{238}U$ age of 594 ± 10 Ma with the youngest population clustering at ca. 600 Ma. One concordant grain was dated at 323 ± 5 Ma, but this grain is suspected to be a contaminant inserted during sample preparation. Of the total dated zircon grains, 51% show Neoproterozoic ages ranging from 594 to 855 Ma. Late Mesoproterozoic zircon grains are less abundant, spanning from 1046 to 1215 Ma. Paleoproterozoic zircon grains are also rather abundant, comprising 30% of the dated zircons with a main peak at ca. 2133 Ma. One Archaean zircon grain was analysed yielding a concordant $^{207}Pb/^{206}Pb$ age of 3231 ± 24 Ma (Fig. 6A).

Sample DKE-425 was collected in the Dahou-Mahou Basin close to the locality of Savalou and consists of a polymictic matrix-supported conglomerate. This sample was collected close to the porphyritic dacite dated at 545.7 ± 7.8 Ma (sample DKE-427) that constrains the minimum age for the deposition of the conglomerate. The youngest zircon grain found in the conglomerate yielded a $^{206}Pb/^{238}U$ age of 563 ± 7 Ma with the youngest population clustering at ca. 585 Ma. The Neoproterozoic zircons grains (n = 52) predominate with 85% of the total reliable ages, ranging from 563 to 705 Ma. Only few Paleoproterozoic zircons grains (n = 3) were found in this rock sample. A small peak is defined at ca. 2500 Ma and the oldest Archaean (n = 6) grain yielded a $^{207}Pb/^{206}Pb$ age of 3301 ± 40 Ma (Fig. 6H).

5. Discussion: tectonic implications and Brazilian connections

5.1. Basement of the Benino-Nigerian Shield

Previous geochronological studies have already indicated that the ages of basement rocks of the Benino-Nigerian Shield range from 2.0 to 2.2 Ga (e.g. Affaton et al., 1978; Caen-Vachette et al., 1979; Agyei et al., 1987; Agbossoumondé et al., 2007; Kalsbeek et al., 2012; Attoh et al., 2013). With the exception of sample DKE-369 (2094 ± 7 Ma) of the Kara leucogneiss, which had already been dated by Kalsbeek et al. (2012) and which we knew to be old, we expected all other granitic samples to be Neoproterozoic. Contrary to our expectations two other samples yielded Paleoproterozoic ages. The granulitic granodioritic gneiss (sample DKE-410) collected in the vicinity of the Kandi Lineament yielded an age of 2085 ± 14 that is within error of the crystallization age of the Kara leucogneiss. Additionally, the schollen from the anatectic gneiss (sample DKE-384) yielded an age of 2153 ± 7 Ma that can be correlated to the Palimé-Amlamé plutonic complex in southern Togo dated at 2127 ± 2 Ma (Agbossoumondé et al., 2007).

In NE Brazil, granulitic gneisses of the Cariré Region of the Ceará Central Domain also occur along the vicinity of the correlated Transbrasiliano Lineament and have similar ages (Amaral et al., 2012) that would suggest a direct correlation to the granulitic gneiss of Benin.

Agbossoumondé et al. (2007) argues that the Benino-Nigerian Shield is part of the WAC, rifted away from it during the early Neoproterozoic. However, due to the occurrence of the major collisional suture zone in the external zone of the Dahomey belt, represented by the alignment of eclogitic mafic massifs (Fig. 2), it seems reasonable to correlate the basement of the Benino-Nigerian Shield with the basement exposed further east that crops out in Nigeria, rather than the basement related to the WAC to the west. Accordingly, in the collisional context, the Benino-Nigerian Shield should be regarded as the upper plate and the down-going easterly dipping WAC as the lower plate (see Fig. 10). The geological link of the Benino-Nigerian Shield and the old basement of Borborema Province with other large cratonic areas remains elusive and should be a topic of further investigation.

5.2. Timing and sources of arc-related and post-collisional magmatism in the Benino-Nigerian Shield

Present zircon U–Pb geochronological data indicate that magmatism related to plate convergence in the active margin of the Dahomey belt spanned from ca. 670 to 545 Ma (Kalsbeek et al., 2012 and this study) and is geographically concentrated in the Benino-Nigerian Shield, the overriding plate. Recently, the timing of continental collision has been constrained to 609 ± 6 Ma by the age of UHP metamorphism of eclogites from the passive margin of the WAC (e.g. Atacora Structural Unit) subducted to mantle depths (Ganade de Araujo et al., 2014a). Such a scenario is only possible during the proposed east-dipping continental subduction that would carry the flanking passive margin to >90 km and ultimately leading to the continental collision of the WAC (lower plate) and the basement of the Benino-Nigerian Shield (upper plate). This implies that granitoids crystallized between 670 and 610 Ma are likely related to the consumption of the Goiás-Pharusian oceanic plate and the hornblende-biotite granodioritic gneiss (sample DKE-419) dated at 610.6 ± 9.4 Ma represents the latest igneous rocks related to arc magmatism. Granitoids younger than ~610 Ma, post-date collision, and are unrelated to subduction zones.

Geochemical data and Sr–Nd isotopic systematics applied to the 670–610 Ma granitoids (now mostly gneisses and migmatites) of the Benino-Nigerian Shield suggested a continental arc affiliation for these rocks (Affaton, 1990; Kalsbeek et al., 2012). This is broadly confirmed here by Hf isotopic systematics, with the exception of the metatonalite sample DKE-358 dated at 643.7 ± 9.8 Ma, which yielded positive $\varepsilon Hf_{(t)}$ values between +7.2 and +10.8, and is interpreted to be juvenile. All other granitoids in the range of 656–610 Ma, yielded negative $\varepsilon Hf_{(t)}$ (−27.1 to −6.4), indicating predominant contribution of older crustal material in their petrogenesis. Our zircon $\varepsilon Hf_{(t)}$ is in general agreement with the whole-rock $\varepsilon Nd_{(t)}$ for the granitoid rocks from Benino-Nigerian Shield which ranges from −19 to −2.9 with two groups of Nd model ages of 1.5–2.1 Ga and 2.2–2.6 Ga, respectively (Kalsbeek et al., 2012). Based on coupled Nd–Sr analyses, Kalsbeek et al. (2012) suggest that these isotopic values are due to mixing of juvenile mantle-derived melts with older continental crust.

In the Borborema Province in NE Brazil, the Santa Quitéria arc evolved from an early juvenile environment into a mature continental arc, which was active during an equivalent time period to those recorded in the Benino-Nigerian Shield (Fig. 7). The juvenile arc magmatism started at ca. 880–800 Ma and continued to 650 Ma as shown indirectly by the Hf values of detrital zircon grains from syn-orogenic deposits. The more mature arc period from ca. 660 to 630 Ma is characterized by hybrid mantle–crustal magmatic rocks (Ganade de Araujo et al., 2012a, 2014c). In contrast with the Benino-Nigerian Shield, in NE Brazil the arc granitoids were extensively reworked at ca. 620–610 Ma forming large areas of diatexites (Ganade de Araujo et al., 2014c).

The Kabye Massif, which occurs within the suture zone and is thrusted over the Atacora passive margin rock sequence (Fig. 2), is thought to represent the deep-seated roots of a continental arc (Duclaux et al., 2006). The massif consists of mantle-derived rocks with cumulates of gabbros and pyroxenites (now HP granulites) that are fractionated into tonalites towards their upper section (Duclaux et al., 2006). U–Pb zircon geochronology for the differentiated tonalites yielded a crystallization age of 623 ± 5 Ma (Ganade et al., unpublished), which can be correlated with the granodioritic gneisses (samples DKE-390 and 413). We therefore also interpret the mafic rocks of the Kabye Massif to represent a bottom section

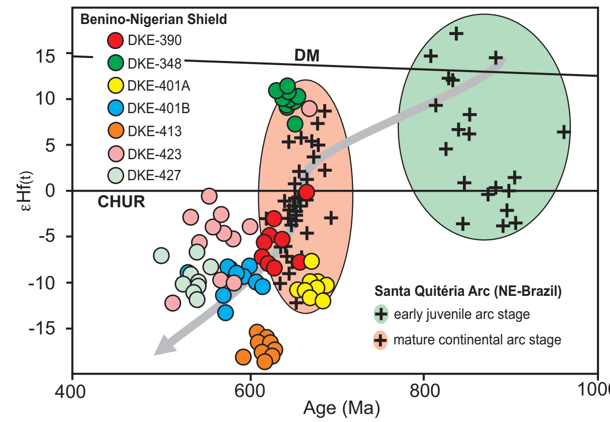

Fig. 7. Diagram for Lu–Hf isotopic evolution vs. U–Pb age for zircon crystals from protolith of granitoids and migmatites from the Benino-Nigerian Shield.
Source: Data from Santa Quitéria arc from Ganade de Araujo et al. (2014c).

of the continental arc emplaced tectonically into shallower crustal levels as a result of continental collision.

In the Benino-Nigerian Shield all igneous rocks dated so far are younger than ca. 670 Ma, however detrital zircon grains from orogenic sedimentary deposits indicate that Neoproterozoic magmatism can be as old as 780 Ma (Fig. 8B). This early stage of magmatism is well described along the West Gondwana Orogen in Africa where it has been ascribed to a juvenile environment based on Nd isotopes and bulk-rock geochemistry (Berger et al., 2011; Caby, 1989; Dostal et al., 1994). In Central Brazil a similar juvenile arc has also been documented (Laux et al., 2005; Pimentel and Fuck, 1992; Pimentel et al., 2000; Matteini et al., 2010) and indicates that part of the Neoproterozoic growth of the 4000-km-long West Gondwana Orogen occurred firstly during the Tonian and Cryogenian periods (950–750 Ma). However, precise paleogeographic reconstructions of these arc systems and the nature of their settings (intraoceanic vs. continental arcs) still need to be better understood.

5.3. Timing of continental collision and crustal thickening

In younger collisional orogens, eclogite facies metamorphism, including UHP rocks is one of the best markers of the onset of the collisional process (e.g. de Sigoyer et al., 2000; Gilotti, 2013; Leech et al., 2005; Liou et al., 2004) and as expressed earlier, timing of collision is well constrained at 609 ± 6 Ma in the Dahomey belt (Ganade de Araujo et al., 2014a).

Zircon ages from the diatexite of sample DKE-423 suggest that partial melting recorded in this sample started at 567 ± 8 Ma remaining hot enough to yield melt-precipitated zircon rims at 560 ± 6 Ma (Fig. 5K). Constraints on the timing of crustal melting in the Benino-Nigerian Shield also come from the felsic dyke represented by our sample DKE-401B dated at 572 ± 8 Ma and from melt-precipitated zircon crystals of one anatectic granite dated at 577 ± 15 by Kalsbeek et al. (2012). Accordingly, the main time of crustal melting and generation of the migmatites in the Benino-Nigerian Shield is 30 to 50 m.y. after the beginning of the continental collision and UHP metamorphism. However, and in spite of the large error associated with the zircon rims of the migmatite sample DKE-387, it suggests that syn-collisional partial melting could have started as early as ca. 585 Ma.

In the Himalayas, ages for the coesite-bearing UHP eclogites are 45–55 Ma (de Sigoyer et al., 2000; Donaldson et al., 2013; Kaneko et al., 2003) while geochronology indicates that peak metamorphism and crustal partial melting took place around 25–20 Ma (e.g. Harrison et al., 1998), a gap of 30–20 m.y. between the beginning

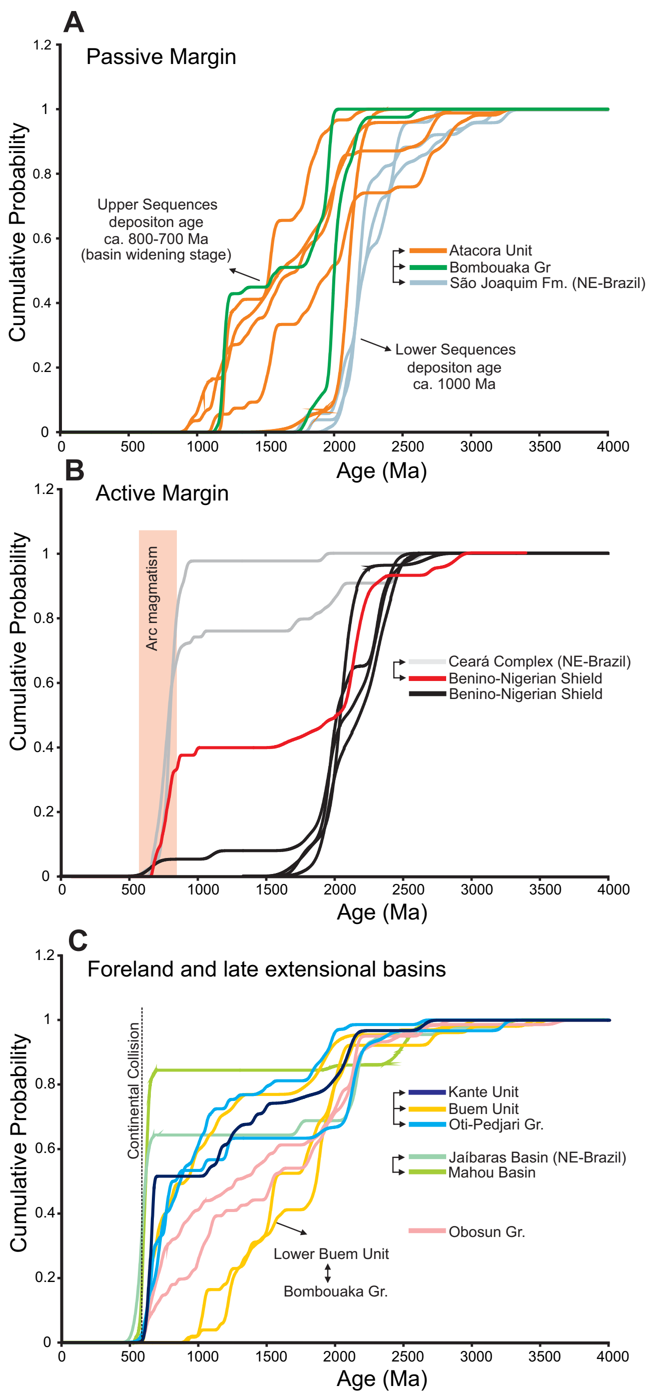

Fig. 8. Cumulative probability plot for the U–Pb ages from supracrustal rocks of the Dahomey belt, Benino-Nigerian Shield and NE Brazil according to the proposed tectonic settings. (A) Passive margin. (B) Active margin. (C) Foreland and late Cambrian basins.

Source: Data from: Kalsbeek et al. (2008, 2012) and Ganade de Araujo et al. (2012b) and from this study.

of collision and the main period of crustal melting. There is also considerable evidence in the form of relict metamorphic minerals, and isotopic ages indicating earlier metamorphism at around 35–30 Ma (e.g. see Hodges, 2000). Thus, widespread melting of mid-crustal levels is thought to have started at around 30 Ma, with more voluminous magmatism at around 20 Ma, and with melt present today ~15 km below the surface underneath the Tibetan Plateau (Unsworth et al., 2005). The time between the UHP metamorphism and the most obvious melting event in the Dahomey belt is therefore similar to that observed in the Himalayan Orogen. Such temporal relations seem to be in agreement with thermal modelling of crustal thickening, which suggests that fertile crust thickened to 65–70 km, in continental collisional settings, would produce copious granite magmatism only after 30–40 m.y. if its average heat-production is >1.2 μ W m^{-3} (Bea, 2012).

These results contrast with findings in NE Brazil, where anatexis of the Tamboril Santa Quitéria Complex associated with the generation of abundant diatexite sectors (Tamboril unit) took place at ca. 620–615 Ma, roughly at the time of collision (Ganade de Araujo et al., 2014c). However, ages of melt-precipitated zircon rims at ca. 580 Ma (Ganade de Araujo et al., 2012b; Santos et al., 2014) from sedimentary-derived migmatites (forearc deposits) adjacent to the Tamboril-Santa Quitéria Complex also suggest that melting related to crustal thickening occurred 30–40 m.y. after UHP metamorphism and initial collision in NE Brazil.

5.4. U–Pb detrital zircon ages

5.4.1. Passive margin

The correlations of the undeformed sedimentary rocks of the Volta basin with the strongly deformed rocks of the Buem and Atacora structural units have long been proposed (Grant, 1967, 1969; Bozhko, 1969; Saunders, 1970; Bondesen, 1972). Observations from the Afram valley (today flooded by the Volta Lake) suggested gradual transitions between the Oti-Pedjari Supergroup and the Buem Structural Unit (Grant, 1967) and between the Bombouaka Supergroup and Atacora Structural Unit (Saunders, 1970; Bondesen, 1972). Earlier detrital zircon studies also suggested a strong correlation between the provenance patterns of the Bombouaka Supergroup and Atacora Structural Unit (Kalsbeek et al., 2008), corroborating original geological observations and correlations.

Two different provenance patterns can be observed by comparing the quartzites of the Atacora Group (Fig. 8A). Sample DKE-353, which is the host of the UHP eclogites of the Lato Hills, has the youngest zircon at ca. 930 Ma, with age distribution peaks at 2200–1700 Ma, 1500–1600 Ma and 1000–1200 Ma, similar to the distribution peaks of zircon ages of the quartzites of the Atacora Structural Unit analysed by Kalsbeek et al. (2008). On the other hand, the youngest zircon from the quartzite of sample DKE-367 from the Atacora quartzitic rocks is ca. 1750 Ma old, with age distributions dominated by Paleoproterozoic zircon grains. This provenance pattern is similar to that of quartzites from the Martinópole Group (São Joaquim Fm.) in NE Brazil, taken as a direct correlative to the Atacora Group (Santos et al., 2008; Ganade de Araujo et al., 2012b). The Bombouaka Supergroup of the Volta Basin (Kalsbeek et al., 2008) also has the same two different patterns found in the Atacora and Martinópole groups (Fig. 8A). Therefore, provenance patterns among these three units indicate that they shared the same source area. However, other evidence suggests that their ages of deposition are different. An Rb–Sr isochron age of 959 ± 62 Ma on clay mineral fractions from the lower units of the Volta Basin suggests that the Bombouaka Group may have been deposited at ca. 1000 Ma (Clauer, 1976). However, the age of the mafic basaltic protolith of the UHP eclogites found in the Atacora Structural Unit yielded an age of 703 ± 8 Ma (Ganade de Araujo et al., 2014a), whereas volcanic rocks interleaved with the São Joaquim quartzites yielded an age of 777 ± 11 Ma (Fetter et al., 2003). A possible cause for the different patterns and age of deposition of the lower units of the Volta Basin and related deformed quartzites may derive from the intrinsic long evolution of the continental passive margin where the sediments have been deposited. In this case, the lower stratigraphic units of the Bombouaka Supergroup being deposited closely in time to the initial opening of the basin (ca. 1000 Ma) with detrital zircon distribution dominated by older ages. The following 250 m.y. of basin widening and magmatic activity allowed younger detrital zircon grains (e.g. 1000–950 Ma) to be incorporated in the upper sections of the detrital record and thus yielding the pattern observed in Fig. 8A and exemplified by the sample DKE-353. Further detailed investigations on the detrital zircon responses related to specific volcanic layers and controlled stratigraphic positions are needed to tackle better this issue.

5.4.2. Active margin

As discussed above the time span from 670 to 610 Ma represents the timing of emplacement of arc-related granitoids in the Benino-Nigerian Shield. The youngest detrital zircon dated at 617 ± 9 Ma indicates that the sedimentary protolith of sample DKE-387, collected in the Benino-Nigerian Shield, was deposited during arc magmatism. Neoproterozoic zircon grains in this sample vary from 781 to 617 Ma with a pronounced peak at ca. 690 Ma, which suggests that arc-related rocks could be as old as ca. 780 Ma. Detrital zircon grains from sedimentary-derived migmatites exposed in the west of the Santa Quitéria Arc, in NE Brazil, yielded similar Neoproterozoic ages from 850 to 640 Ma (Fig. 8B) and were interpreted as forearc sequences with most of the detritus originating from the erosion of arc rocks exposed to the east (Ganade de Araujo et al., 2012b). Older Early Neoproterozoic igneous rocks have not yet been found in the Dahomey belt, however correlations along strike in the West Gondwana Orogen indicate that magmatism related to the early stages of the subduction in the Pharusian-Goiás Ocean can be as old as 800–900 Ma (Berger et al., 2011; Cordani et al., 2013a,b; Ganade de Araujo et al., 2014c).

5.4.3. Foreland basin

Affaton (1990) proposed that the Buem Structural Unit should be divided into two successions separated by glaciogenic deposits and suggested that the lower and upper portions of this unit should be correlated to the Bombouaka and Oti-Pedjari Supergroups, respectively. Indeed, our sample DKE-361 from the upper portion of the Buem Structural Unit shares the same source areas of sample DKE-362 from the Oti-Pendjari Supergroup, suggesting a correlation between these two units (Fig. 8C). Similarly, samples from the Buem Structural Unit investigated by Kalsbeek et al. (2008) also relates to the Bombouaka Supergroup, suggesting an overall correlation between these two successions (Fig. 8C). The schists from the Kante Unit have the same provenance pattern of those observed in the upper Buem Unit and Oti-Pendjari Supergroup, suggesting that these sedimentary sequences shared the same source areas (Fig. 8C). Apart from the lower Buem Structural Unit, which can be correlated to the passive margin system discussed above, all the upper Buem and Kante units and the Oti-Pendjari Supergroup have their youngest detrital zircon grain dated at ca. 600 Ma with a dominant Neoproterozoic (600–900 Ma) age distribution (Fig. 8C). These Neoproterozoic zircon grains were likely shed by the metamorphic and magmatic rocks from the Benino-Nigerian Shield after the continental subduction and collision at ca. 610 Ma. We therefore envisage a foreland basin developed coeval with the building of the collisional chain to the east, with sedimentation associated with the progression of the thrust front.

The youngest zircon grains found in the Obosum Group are even younger than the lower sequences of the Volta basin with ages at ca. 580 Ma (Kalsbeek et al., 2008), which indicates continual infilling of the foreland basin during thrusting related to melting-assisted exhumation within the Dahomey belt.

5.4.4. Late extensional Cambrian basins

After crustal thickening and orogenesis a series of intracontinental extensional basins developed over the older basement or the recently created Neoproterozoic crust in the Benino-Nigerian Shield and Borborema Province in NE Brazil. In Benin, timing of Dahou-Mahou Basin formation can be constrained by the 545.7 ± 7.8 Ma age of volcanism (sample DKE-427). Detrital zircon provenance of related conglomeratic layers in the Dahou-Mahou Basin revealed a predominance of late Neoproterozoic zircon grains with the youngest at 563 ± 7 Ma. In NE Brazil, the Jaibaras basin, along the axis of the Transbrasiliano-Kandi Lineament, consists of a basal fault-scarp-related sedimentary continental package (Oliveira and Mohriak, 2003). There, volcanism is dated at 537 ± 8 Ma (Garcia et al., 2010) and detrital zircon ages (Ganade de Araujo et al., 2012b) are remarkably similar to those found in Dahou-Mahou basin (Fig. 8C). Therefore, extensional basins developed throughout the orogen between 550 and 540 Ma, and mark the end of thrusting and the beginning of extension along the strike-slip collisional shear zones.

5.5. The role of the Transbrasiliano-Kandi Lineament in the final orogenic configuration

There is an agreement in the literature that the subvertical Transbrasiliano Lineament in South America corresponds to the Kandi Lineament in Africa, thus providing a key constraint on the present-day fit between these continents (Caby, 1989; De Wit et al., 2008a,b). However, Kalsbeek et al. (2012) proposed that the Transbrasiliano Lineament should be correlated with a shear zone 150 km to the east of the Kandi Lineament in Benin. According to this fit, the Neoproterozoic granitic rocks of the Benino-Nigerian Shield would correlate with Neoproterozoic granitic rocks ~150 km to the west of the Transbrasiliano Lineament in NE Brazil. This correlation is not supported by lithological evidence since equivalent Neoproterozoic granitic rocks are missing to the west of the Transbrasiliano Lineament, which is in fact dominated by post-collisional granitoids of ca. 560–530 Ma. Additionally, reliable plate reconstructions using marine magnetic anomalies, ocean floor topography and homologous continental and oceanic structures suggest a reasonable fit between the Transbrasiliano and Kandi Lineaments (e.g. Moulin et al., 2010 and references therein).

According to this fit, the projection of the Atacora Structural Unit into NE Brazil should follow a southwesterly trend (Fig. 9), following the documented high-density gravity anomalies beneath the Parnaíba Basin (Lesquer et al., 1984). In this case, the proposed fit precludes a direct along-strike correlation between the Martinópole Group and Atacora Structural Unit, as proposed here and by others (e.g. Caby, 1989; Santos et al., 2008). In NE Brazil, the Martinópole Group crops out immediately west of the Transbrasiliano Lineament, 200 km eastward from the projected continuation of Atacora Structural Unit (see Fig. 9). Several reasons could explain this misfit including: (i) loss of the rock record (width of crust) during crustal extension and rifting in the Equatorial Atlantic; (ii) Neoproterozoic continental collision deformation responsible for large-scale tectonic transport, shortening and dismembering of crustal masses and lithotectonic assemblages; (iii) extent of the passive margin deposits; and (iv) and most likely a combination of the above processes.

The crustal structure of the Parnaíba Block (a key block of Precambrian crust concealed under the Parnaíba Basin) revealed by airborne gravity and magnetic data (Castro et al., 2014) demonstrates that supracrustal rocks of Martinópole Group could be correlated with rocks of the Gurupi Belt located further west; an idea already put forward by Klein and Moura (2008). On the other hand, the sub-vertical shear zones of the Transbrasiliano Lineament covered by the basin control the eastern limit of the geophysically inferred Parnaíba Block. New deep seismic reflection data along the Parnaíba Basin revealed that the sub-vertical shear zones of the Transbrasiliano Lineament truncate moderate (45°) easterly dipping reflectors to the east of the lineament (Daly et al., 2014). These reflectors reach the top of the mantle and could be the evidence of

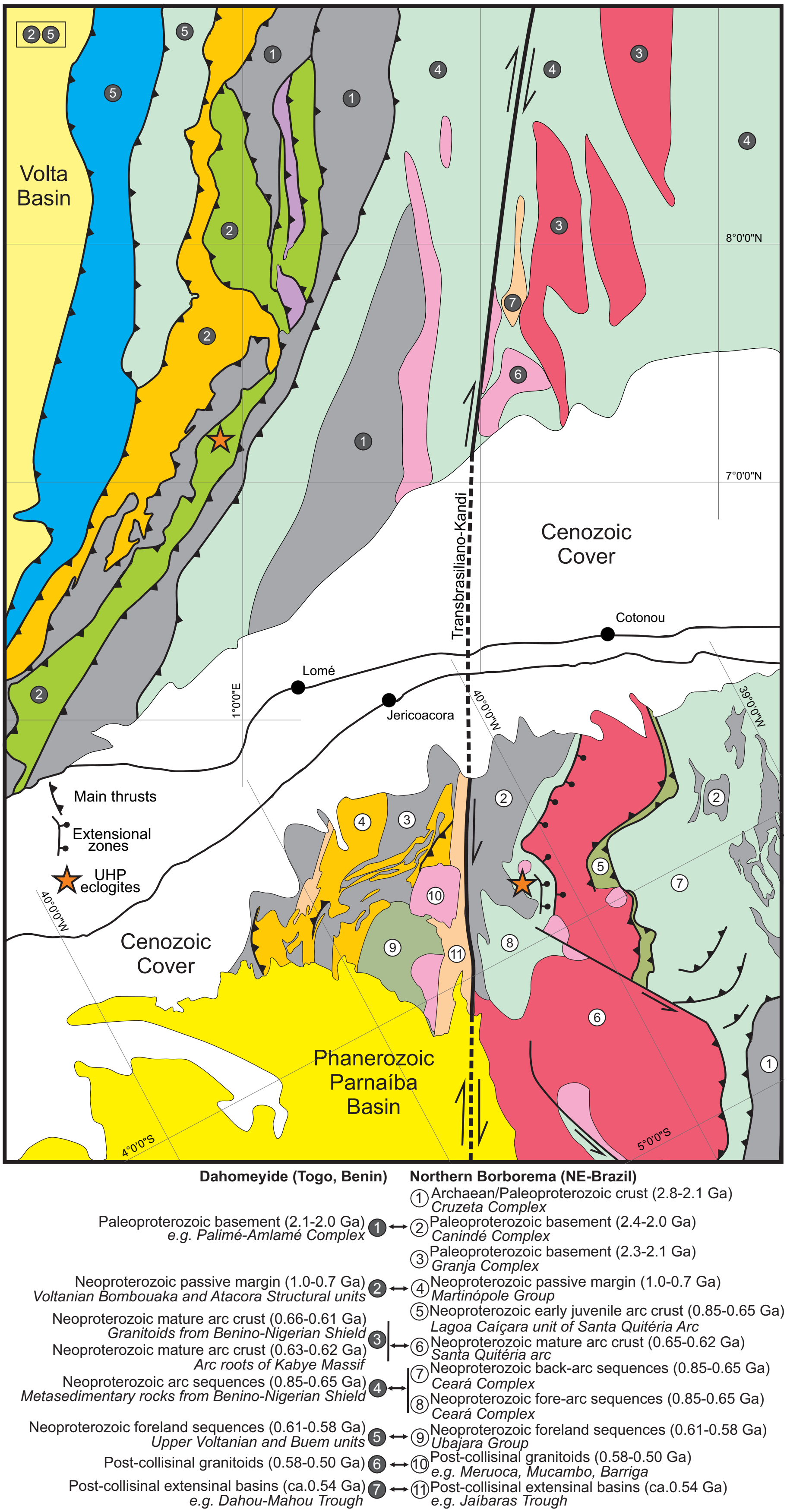

Fig. 9. Pre-drift fit between the Dahomey belt and northern Borborema Province using the Transbrasiliano-Kandi Lineament as the main tie element.

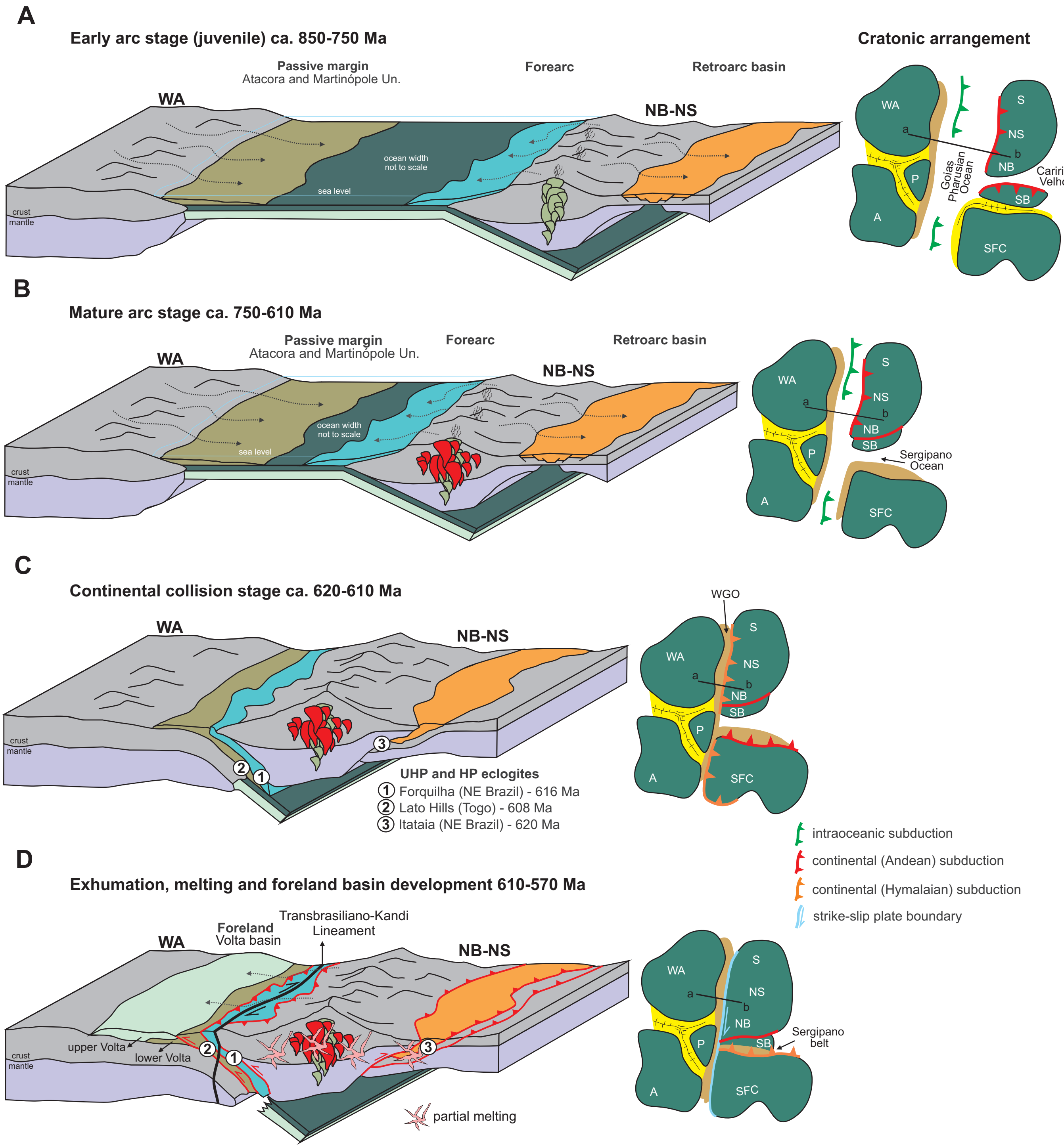

Fig. 10. Proposed tectonic evolution of the Dahomey belt and correlative Northern Borborema Province with evolution of the cratonic arrangement throughout time. (A) Development of West Africa craton (WA) flanking passive margin from 1000 to 700 Ma and concomitant development of early continental arc assemblages (ca. 800 Ma) in NE Brazil over the basement of Northern Borborema Province (NB) and Benino-Nigerian Shield (NS). Cratonic arrangement illustrates the conjoined West Africa-Parnaíba-Amazonian cratonic block with intervening intracontinental and/or restricted oceanic basins; intraoceanic arc systems of the Goiás-Pharusian Ocean (e.g. Pimentel and Fuck, 1992; Berger et al., 2011); the 1000–960 Ma Cariris Velhos continental arc (Santos et al., 2010) and the development of the Sergipano basin between the northern São Francisco craton and Southern Borborema Province (present day reference). (B) Subductions of the oceanic lithosphere and development of the mature arcs in the Benino-Nigerian and Northern Borborema Province. Cratonic arrangement illustrates the docking of the Northern Borborema against the Central Borborema (Cariris Velhos) and development of the Sergipano passive margin. (C) Continental collision and formation of the UHP terranes in Togo (passive margin subduction) and NE Brazil (fore arc subduction) and inversion and incipient subduction of the back-arc basin of Santa Quitéria arc. Cratonic arrangement illustrates the suturing of the main cratonic blocks along the West Gondwana Orogen and the continental arc related to the closure of the Sergipano Ocean. (D) Exhumation of UHP terranes and melting related with crustal thickening. Cratonic arrangement illustrates the collision of the Southern Borborema Province and the Northern São Francisco craton at 580 Ma and the development of the earlier (ca. 610 Ma) strike-slip transference zone (transformant plate boundary).

the fossil east-dipping Neoproterozoic subduction in West Gondwana Orogen, recorded in the Borborema Province. The obducted transitional passive margin deposits of the Novo Oriente basin (e.g. Ganade de Araujo et al., 2010) could therefore represent the continuation further south of the large passive margin system (Atacora Structural Unit) discussed earlier (see Fig. 1).

The dextral Transbrasiliano-Kandi strike-slip belt formed as a result of relative movement obliquity and escape between the continental blocks involved in the collision along the West Gondwana Orogen at ca. 610 Ma (e.g. Caby, 1989; Castaing et al., 1994; Ganade de Araujo et al., 2014a,b) (Fig. 10). This strike-slip belt reactivated the previous sutures and displaced lithotectonic assemblages and former neighbouring terranes and units. According to Ganade de Araujo et al. (2014b), the lineament was active as a transform plate boundary, which allowed the approximation of the Borborema Province to the São Francisco craton, leading to closure of an oceanic tract and final collision between these two continental blocks at ca. 590 Ma.

Assuming that this strike slip system post-dated collision at 610 Ma, the 20 m.y. elapsed from the first nucleation of shear zone system during collision in the WGO (ca. 610 Ma) to its movement cessation after the São Francisco craton collision (ca. 590 Ma) could be used to quantify the displacement along the transform boundary (Fig. 10). Assuming ordinary plate velocities of 50 mm y^{-1} would result in a displacement of 1000 km. Accordingly, conservative (20 mm y^{-1}) or faster (70 mm y^{-1}) plate velocities would result in smaller and larger displacements of 400 and 1400 km, respectively. Consequently, taking ordinary plate velocities and assuming that the width of crust lost during Atlantic rifting was of the order of 50–100 km (Reeves et al., 2002) and discarding the effect of subsequent reactivations, the initial position (before collision in the WGO) of the displaced lithotectonic assemblages would be 900–1100 km from the present day position. This would have major implications to the fit we see today. For example, before collision, the Santa Quitéria arc and its basement would occupy the region of the present day Benino-Nigerian Shield and continental collision of the Northern Borborema basement block would be against the WAC, instead of the Parnaíba Block (Fig. 10).

This scenario also has major implications for the correlation of the UHP eclogites of the Dahomey and those of the northern Borborema Province. One of the misfits put forward by many authors about the correlation of these rocks is that they crop out on different sides of the Transbrasiliano-Kandi Lineament (Santos et al., 2008; Kalsbeek et al., 2012). However, one has to bear in mind that the tectonic settings of these eclogites are different. The UHP eclogites of the Lato Hills, in Togo are found within the metasedimentary rocks that make the passive margin system of the subducting WAC, whereas those from NE Brazil are part of the overriding plate are were juxtaposed with fore-arc strata during continental subduction (Fig. 10). In this way, restoring the displacement of the Transbrasiliano-Kandi Lineament would place both UHP eclogites along the same line across the orogen. Thus, when continental collision started, the first lithotectonic unit to be subducted should be the remaining fore-arc strata followed by the progressive subduction of the WAC passive margin. This would explain the slightly older age for the UHP metamorphism in the NE Brazil eclogites (616 ± 4 Ma) when compared to the age of UHP metamorphism in the subducted passive margin in Togo (608 ± 6 Ma) (Ganade de Araujo et al., 2014a).

6. Conclusions

The Dahomey belt preserves a well-organized orogenic architecture developed during the Neoproterozoic. Our new zircon isotopic data added to previously acquired data from the Borborema Province (NE Brazil) allow the following conclusions.

- Passive margin deposits of the Atacora Structural Unit and lower units of the Volta Basin have detrital zircon signatures compatible with the flanking West Africa Craton. This signature is similar to that of the Martinópole Group in NE Brazil, taken as a direct correlative of the Atacora Structural Unit.
- Arc-related magmatism originated due to the east-dipping subduction of the Goiás-Pharusian oceanic lithosphere is represented by a variety of granitoids emplaced in the Benino-Nigerian Shield between 670 and 610 Ma. These granitoids were mainly sourced from crustal reservoirs with subordinate juvenile input. Detrital zircon grains from syn-orogenic deposits in Benino-Nigerian Shield suggest that arc development could have started as early as 780 Ma.
- The main period of melting related to crustal thickening appears only ca. 30 m.y. after initiation of the continental collisional event marked by the ca. 610 Ma UHP metamorphism.
- Foreland development represented by the upper units of the Volta basin developed soon after continental collision and persisted with the development of the west-verging thrust front synchronously with the main period of melting at 580 Ma.
- Restoration of the movement of the Transbrasiliano-Kandi Lineament would place the Dahomey belt and Borborema Province along the same section of the West Gondwana Orogen.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.precamres.2016.01.032.

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