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Highlights

• Zircon U-Pb-Hf isotope data suggest mainly juvenile growth between 2.3–2.1 Ga
• Reworking of Archaean crust in southern Ghana is confined to between 2.141–2.126 Ga
• Combined isotope data suggest subduction related crustal growth
• Emplacement of 2.23 Ga granodiorite contradict suggested plume initiated subduction
• An evolutionary model is proposed
Zircon U-Pb-Hf evidence for subduction related crustal growth and reworking of Archaean crust within the Palaeoproterozoic Birimian terrane, West African Craton, SE Ghana

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Abstract

Zircon Lu-Hf isotopic data from granites of southern and northwestern Ghana have been used to investigate the contribution of reworked Archaean bedrock to the Birimian crust of Ghana, West African Craton. Zircon from seven localities in southern Ghana and one locality in western Ghana were analysed. Combined U-Pb and Lu-Hf isotope data suggest juvenile crustal addition between 2.3–2.1 Ga, with a short period of reworking of Archaean crust. Until now, evidence for reworking of Archaean basement during Birimian magmatism in Ghana has hinged on whole-rock Nd model-
ages of the Winneba pluton, and sparse inherited zircon grains from mainly northwestern Ghana.

Our data suggest that reworking of Archaean crust is greater than previously inferred, but was limited to between ~2.14–2.13 Ga. This period of reworking of older crustal components was preceded and succeeded by juvenile crustal addition.

Coupled isotopic data suggest an eastward, mainly retreating arc system with a shorter pulse of accretion between ~2.18–2.13 Ga and a rapid return to slab retreat during the growth of the Birimian terrane. The accretionary phase initiated melting of sub-continental lithospheric mantle and the overlying Archaean crust, generating magma with sub-chondritic Hf signatures. Subsequent slab retreat led to trench-ward movement of the magmatic activity and the mixture of juvenile and Archaean crust was replaced by uncontaminated juvenile magma.

The 2.23 Ga age of the West Accra granodiorite (PK105) demonstrates the emplacement of felsic rocks during the Eoeburnean and pre-dates the suggested plume related rocks, contradicting suggested plume initiated subduction.

1. Introduction

The formation of the Birimian terranes in West Africa (Fig. 1a–c) occurs towards the end of a period sometimes assumed to be associated with global magmatic quiescence (Condie, 2009). Yet, the formation of the Birimian crust has been cited as an example of rapid crustal growth, as large volumes of juvenile continental material were emplaced during a short time-span (Abouchami et al., 1990). Crystallisation ages from the Birimian bedrock of Ghana range between ~2.31 and 2.06 Ga, with a predominance of ages between 2.21 and 2.06 Ga (e.g. Gasquet et al., 2003; de Kock et al., 2011). These rocks have largely juvenile Nd isotope signatures (Abouchami et al., 1990; Liégeois et al., 1991; Boher et al., 1992; Ama-Salah et al., 1996; Hirdes et al., 1996; Doumbia et al., 1998; Gasquet et al., 2003; Pawlig et al., 2006; Klein et al., 2008; Tapsoba et al., 2013) with the exception of the Winneba pluton from southeastern Ghana, which has a $\varepsilon_{\text{Nd}}(2.173 \text{ Ga}) = -5.3$ and a depleted
mantle model age of ~2.6 Ga, indicating the involvement of Archaean crust (Taylor et al., 1990; Leube et al., 1990). Based on trace element geochemistry of mafic metavolcanic rocks, it has been proposed (Abouchami et al., 1990) that the Birmian crust formed rapidly and in response to mantle plume activity. Although alternative views such as arc accretion and convergent magmatism have been proposed (e.g. Sylvester and Attoh 1992; Feybesse and Milési 1994; Ama-Salah et al. 1996; Pouclet et al. 2006; Baratoux et al. 2011; de Kock et al. 2012), the Birimian terranes are still widely promoted as a prime example of mantle plume-related crust formation (c.f. Arndt, 2013).

Feybesse et al. (2006) propose that the onset of continental crust growth within the Birimian terrane started at the end of the Eoeburnean (c. 2.35–2.15 Ga) phase, with the intrusion of abundant monzogranites between 2.16–2.15 Ga. Reworked Palaeoproterozoic to Archaean crust within the Birimian terrane is, apart from the Winneba pluton in southeastern Ghana (i.e. near the SE margin of currently exposed Birimian crust), only known through the presence of 2.26–2.88 Ga xenocrystic and commonly discordant zircon from rocks in the Bolé-Navrongo belt in northwestern Ghana e.g. the Gondo orthogneiss and the Ifantayire granite gneiss (Thomas et al., 2009; Siegfried et al., 2009; de Kock et al., 2011, Fig. 1c). Available geochronological data for the Birimian terrane, whole rock Lu-Hf and Sm-Nd isochrons for basalts (Blichert-Toft et al., 1999) and zircon U-Pb of granites (Hirdes et al., 1996) are coeval within error, i.e. they formed at 2.15 ±0.05 Ga. Following a similar approach as Næraa et al. (2012), we explore coupled shifts in zircon U-Pb–Lu-Hf isotopes to explore crustal growth and reworking of older crust within an accretionary orogen. Detrital zircon δ18O from five rivers draining Birimian bedrock in Ghana yield a weighted mean of 6.7 ±0.2 (MSWD = 5; Kristinsdóttir, 2013), which might indicate a significant reworked supracrustal component (c.f. Dhuime et al., 2012). Such an inference is in stark contrast with current models for the Birimian continental crust growth, which imply that the entire mass of juvenile crust formed around 2.15 Ga with the exception of the 2.173 +0.107/-0.115 Ga Winneba pluton (Taylor et al., 1988; Leube et al., 1990; Taylor et al., 1992).
As the median Birimian mantle composition as defined by Blichert-Toft et al. (1999) virtually coincides with the new crust curve presented by Dhuime et al. (2011) but is markedly lower than e.g. coeval depleted mantle values proposed by Griffin et al. (2000), the new crust curve of Dhuime et al. (2011), inferred from modern island arc basalts, is used as the depleted mantle reference in further discussion.

The samples in this study are mainly from the southernmost part of Ghana with the exception of the samples from the Sewfi belt granitoid (ASGH022A/C), which are from the Vinson quarry in the mid- to northwestern part of Ghana (Fig. 1). These rocks were sampled with the aim to further investigate the presence of reworked Archaean components within the Birimian terrane.

2. Geological settings

2.1. The southern West African Craton

The Reguibat Shield in the North and the (Leo-) Man Shield in the South make up the West African Craton in NW Africa (Fig. 1a). These Shields are separated by the Neoproterozoic–Palaeozoic Taoudeni basin. Archaean rocks are exposed in the western part of both shields and Palaeoproterozoic rocks of the Baoulé Mossi domain are abundant in the eastern part of the Man Shield (Fig. 1a and 1b). The Baoulé Mossi domain is juxtaposed with the Man Shield and formed along a ~2.1 Ga active accretionary margin during the Birimian event (Sylvester and Attoh, 1992; Feybesse and Milési, 1994; Vidal and Alric, 1994; Ama-Salah et al., 1996; Hirdes and Davies, 2002; Pouclet et al., 2006; Baratoux et al., 2011; de Kock et al., 2012), however, alternative interpretations including formation of continental crust at the margins of an oceanic plateau have been suggested (Abouchami et al., 1990; Boher et al., 1992). The Man Shield and the Baoulé Mossi domain are separated by the Sassandra fault (Abouchami et al., 1990; Attoh and Ekwueme, 1997; Fig. 1b). TTG gneisses > 3.0 Ga make up the oldest component in the Man Shield and are overlain by greenstone belts and in turn intruded by 2.97–2.78 Ga granites (Attoh and Ekwueme, 1997). On
the basis of lithological and age correlation, it has been suggested that the South American São Luis Craton and the Man Shield were united during the emplacement of the Birimian bedrock (e.g. Feybesse et al., 2006).

2.2. Birimian bedrock of the West African Craton

Birimian rocks of the Baoulé Mossi domain consists of 2.25–1.98 Ga volcanic belts, granitic gneisses and sedimentary basins, of which all have been affected by greenschist to amphibolite facies metamorphism (Milési et al., 1989; Boher et al., 1992; Ama-Salah et al., 1996; Hirdes et al., 1996; Peucat et al., 2005; Feybesse et al., 2006; de Kock et al., 2009; Baratoux et al., 2011). Volcanic belts and sedimentary basins trend NE-SW and make up the majority of the Palaeoproterozoic basement of Ghana (Fig. 1c; Leube et al., 1990; Hirdes et al., 1996). The volcanic belts are dominated by tholeiitic basalts at the base and calc-alkaline andesites, dacites and rhyolites in the upper sections (e.g. Boher et al. 1992; Sylvester and Attoh, 1992). The metasedimentary basins are isoclinally folded and consist of volcanoclastic rocks, greywacke, argillitic rocks and chemical sedimentary rocks (Leube et al., 1990). There are four main suites of granite; Winneba, Cape Coast, Dixcove and Bongo. The rocks within the Winneba suite are restricted to a small area near the town of Winneba in southeastern Ghana and occur as granite to granodiorite. These are the only intrusions where Archaean Sm-Nd model ages hint at the involvement of reworked ancient crust (Leube et al., 1990; Taylor et al., 1988; 1992; Fig. 1c). The Cape Coast suite predominantly intrudes the metasedimentary basins and form larger plutons of peraluminous biotite-granodiorites (Leube et al., 1990). Dixcove suite rocks mainly intrude volcanic belts and form smaller plutons of metaluminous hornblende bearing granitic rocks and the younger Bongo type are potassium-rich granitic rocks that are found in northern Ghana and intrude the Tarkwaian sediments (Leube et al., 1990). Granodiorites and tonalities dominate these intrusions and granite (sensu stricto) only account for a minor part (Eisenlohr and Hirdes, 1992). The relative amount of granitic rocks within the volcanic belts in Ghana increase towards the northwest, which
has been interpreted as a function of erosional level, such that northwestern Ghana represents the
deepest crustal sections exposed in the region (Taylor et al., 1992). The events that formed the basin
and belt structure and subsequent geotectonic evolution is termed the Eburnean and prior events are
termed the Eoeburnean (de Kock et al., 2011).

2.3. Growth of Birimian crust in Ghana

The majority of the Birimian terrane within the Baoulé Mossi domain consists of rocks that were
emplaced around 2.2–2.1 Ga (Abouchami et al., 1990; Boher et al., 1992; Ama-Salah et al., 1996;
Doumbia et al., 1998; Gasquet et al., 2003; Pawlig et al., 2006; Klein et al., 2008; Tapsoba et al.,
2013) and with depleted mantle Nd model ages within 300 Myr. of their crystallisation ages (Boher
and Lu-Hf isotopes, the isotopic composition of the Birimian depleted mantle was determined to
$\varepsilon_{\text{Hf}}(2.150 \text{ Ga}) \approx 6 \pm 2$ and $\varepsilon_{\text{Nd}}(2.150 \text{ Ga}) \approx 3 \pm 1$ (Blichert-Toft et al., 1999). The only known exception
with the West African Craton that deviates significantly from this isotopic signature is represented
by granitic rocks found in southeastern Ghana, the Winneba pluton. As noted above, Sm-Nd
isotopic data from this body indicate incorporation of crustal material from an Archaean source
(Leube et al., 1990, Taylor et al. 1992).

Gasquet et al. (2003) recorded a 2.312 ±0.02 Ga (MSWD=8.1 n=8) xenocrystic zircon in a 2.170
±0.019 Ga granite from the Dabakala area (Fig. 1b). These xenocrystic zircon have been suggested
to represent an early phase of crustal growth in the Baoulé Mossi domain. Feybesse et al. (2006)
proposed a geodynamic model where the initial magmatic and tectonic activity that formed the
Birimian bedrock in Ghana started at ~2.35 Ga with deposition of, for example, banded iron
formations, which was followed by extensive emplacement of mafic to ultramafic crustal segments
between 2.25–2.17 Ga. Mafic magmatism was followed by monzogranites between 2.16–2.15 Ga,
which marks the first growth of continental crust in the Birimian terrane (Feybesse et al., 2006).
Metamorphism reached upper greenschist- to amphibolite facies during the Eburnean orogeny between ~2.13–2.00 Ga (Leube et al., 1990; Eisenlohr and Hirdes, 1992; Hirdes and Davis, 1998; Feybesse et al., 2006). Magmatic rocks younger than 2.07 Ga are scarce in the entire Baoulé Mossi domain, indicating a magmatic quiescence after this period (Gueye et al. 2007; de Kock et al. 2011).

3. Samples

3.1. PK101 Amasaman biotite hornblende tonalite (N 05° 42.730'/W 00° 16.270')

This rock is a weakly foliated medium- to coarse-grained biotite hornblende tonalite to granodiorite that was sampled in the central Suhum basin. It is cut by discordant late leucosomes with diffuse contacts with the main rock. It is dominated by nearly equigranular quartz and plagioclase. Antiperthite occurs in some samples. Bioite is usually fresh. Secondary epidote-group minerals and muscovite overgrow feldspar, and minor amounts of intergranular calcite fills pore spaces and fractures (Fig. 2). Zircon is most commonly observed within biotite but also within quartz and feldspar.

The zircon population is euhedral to subhedral and grains vary in size from 50–500µm along their c-axis (Fig. 3). Most grains are oscillatory zoned and BSE-bright, with a thin (<20µm) BSE-dark rim of metamorphic zircon.

3.2. PK102 Nsawam biotite hornblende granite (N 05° 48.660'/W 00° 16.270')

This rock was sampled in a quarry in the town of Nsawam about 60 km northeast of Winneba. It is a coarse grained biotite hornblende granite, with green pleochroic biotite intergrown with abundant hornblende. Euhedral titanite is abundant and defines a weak tectonic fabric with biotite and hornblende. Secondary fine-grained muscovite and epidote overgrows feldspar and medium grained epidote with minor calcite occur along fractures and grain boundaries.
The zircon grains are 100–300μm along their c-axis, euhedral and display distinct oscillatory zonation in BSE (Fig. 4). Thin rims of metamorphic zircon cut the primary zonation in many grains and some grains have a BSE dark metamict appearance along cracks. 

### 3.3. PK103 Gomoa Fetteh hornblende biotite granite (N 05° 26.185’/W 00° 28.372’)

This hornblende biotite granite was sampled in the Krokrobite Tuba quarry close to the coast about 20 km east-northeast of the town of Winneba. Biotite and hornblende are roughly equal in abundance, with a slight tendency for greater amounts of biotite. K-feldspar is more abundant than plagioclase. Perthite is common and myrmekite intergrowths occur. Epidote, sometimes euhedral, is predominantly found along grain boundaries between feldspars and hornblende although some feldspar clouding might be due to fine secondary epidote. Minor amounts of calcite are localised along fractures. The mafic minerals have a slight preferred tectonic orientation.

Zircon grains are subhedral to euhedral and between 50–150 µm along their c-axis (Fig. 5). In BSE, oscillatory zonation is visible in most grains, but grains with a higher abundance of cracks have a more metamict and patchy appearance. Most grains have a thin rim of BSE bright secondary zircon truncating the primary zonation.

### 3.4. PK105 West Accra biotite hornblende granodiorite (N 05° 37.320’/W 00° 19.803’)

This sample is a weakly foliated biotite hornblende granodiorite from the Suhum basin. The rock is coarse grained with patchy occurrence of secondary epidote, muscovite and calcite, mostly along fractures. The feldspar is slightly cloudy due to secondary fine grained epidote or muscovite. Although the rock lacks conspicuous deformation features, quartz has recrystallised into subgrains.

The zircon grains are 50–500µm along their c-axis and morphologically euhedral to subhedral (Fig. 6). The zircon core domains are sector- or oscillatory-zoned any many grains have a thin BSE-bright rim of secondary zircon discordantly cutting zonation in the core.
3.5. ASGH003A Cape coast two-mica granodiorite (N 05° 20.759’/W 00° 36.828’)

The outcrop is located in southern Cape Coast basin and is heterogeneous. Lithologies vary from fine to coarse grained, but medium to coarse grained varieties dominate. Metasedimentary xenoliths have higher contents of mafic minerals, of which biotite dominates. The sample investigated here is a coarse grained muscovite biotite granodiorite. Euhedral muscovite occur in minor amount but the majority is found together with biotite and at grain boundaries. The muscovite is interpreted to be primary. Feldspars are variably altered to sericite and perthite is common. Muscovite sometimes occur as secondary minerals on feldspar. Myrmekite inter-growths occur in minor amounts. The zircon population is between 50–150µm along their c-axis and mostly with euhedral morphology, many with sharp pyramid terminations (Fig. 7). Zircon grains are microstructurally complex, with BSE-bright oscillatory zoned cores discordantly cut by BSE-darker oscillatory zoned rims. Many grains have a patchy, metamict appearance in association with cracks. In less cracked grains the zonation is weaker to almost non detectable.

3.6. ASGH007A Dixcove hornblende tonalite (N 04° 47.609’/W 01° 56.733’)

This hornblende tonalite is intrusive into Birimian volcanic flows and volcaniclastic sedimentary rocks. Angular basalt fragments are common and usually <10 cm in size. Minor amounts of fresh pyrite occur. It is medium to coarse grained rock with recrystallised quartz that forms sub-grains. Feldspars are undeformed and commonly subhedral to euhedral, forming a slightly porphyritic texture. Most grains are strongly saussuritised and sericitised but lack any tectonic fabric. Epidote ranges from fine saussurite to larger grains (100–150µm), and occurs with chlorite and sometimes minor amounts of calcite. The majority of the zircon in this sample is euhedral with sharp pyramid terminations (Fig. 8). In BSE, a weak oscillatory zonation is visible in most grains. The zonation in many grains is more pronounced towards grain boundaries.

3.7. ASGH022A/C Sunyani basin mica granites (N 07° 28.842’/W 02° 11.016’)

These rocks were sampled from the Vision quarry in the Sunyani basin. The rocks within the quarry are diverse, with biotite muscovite to pure muscovite granites that contain schistose metasedimentary xenoliths of varying size (up to tens of metres).

Sample 22A is a biotite muscovite granite, and is the main rock type at the locality. It has abundant primary muscovite and lesser amounts of biotite. Plagioclase is the dominant feldspar but microcline occurs in lesser amounts. Most feldspars are slightly altered, primarily into sericite. The rock is equigranular and recrystallised with many grain boundaries forming 120° triple junctions.

Sample 22C is a late muscovite granitic pegmatite. The main difference between the pegmatite and the two-mica main granite is the near lack of biotite in the former. The feldspar composition is very similar to the two-mica granite (sample 22A), but it is slightly less altered.

The zircon populations in these rocks are identical in terms of morphology and texture and will be described together. Zircon grains are 50–150µm along their c-axis with a subhedral to euhedral morphology. Texturally they vary from well-preserved BSE-bright oscillatory zoned zircon to metamict BSE-dark and patchy zoned zircon (Fig. 9 and 10). Many grains have a thin rim of metamorphic zircon almost always associated with metamict BSE-dark textures.

4. Analytical methods

Zircon separation was done at the Department of Geology, Lund University. Rock samples were crushed by hand on a steel plate and clean chips were pulverised using a Cr-steel swing mill. Heavy minerals were separated on a Wilfley table, and collected in petri dishes. Magnetic fractions were removed using a magnetic pen and zircons were then hand picked under a binocular microscope.

Selected grains were mounted on double sided tape together with the zircon standard 91500 (Wiedenbeck et al., 2004) and cast in epoxy. The epoxy mount was polished to expose internal cross sections through the grains. Back-scattered electron imaging (BSE) was used to investigate internal growth patterns in the individual crystals, and to guide the analytical work.
4.1. Zircon U-Pb dating

Secondary ionisation mass spectrometry (SIMS) U-Th-Pb analyses were carried out using a large geometry Cameca IMS1280 instrument at the Swedish Museum of Natural History in Stockholm. Instrument set up follows that described by Whitehouse et al. (1999), Whitehouse and Kamber (2005) and references therein. An O$_2^-$ primary beam with 23 kV incident energy (-13 kV primary, +10 kV secondary) was used for sputtering. For this study, the primary beam was operated in aperture illumination (Köhler) mode yielding a ca. 15–20 μm spot. Pre-sputtering with a 25 μm raster for 120 seconds, centring of the secondary ion beam in the 3000 μm field aperture (FA), mass calibration optimisation, and optimisation of the secondary beam energy distribution were performed automatically for each run, FA and energy adjustment using the $^{90}$Zr$^{16}$O$^+$ species at nominal mass 196. Mass calibration of all peaks in the mono-collection sequence was performed at the start of each session; within run mass calibration optimisation scanned only those peaks that yield consistently high signals from the zircon matrix, namely $^{90}$Zr$^{16}$O$^+$, $^{94}$Zr$^{16}$O$^+$ (nominal mass 204), $^{177}$HfO$^+$ (nominal mass 209), $^{238}$U$^+$ and $^{238}$U$^{16}$O$^+$, with intermediate peaks adjusted by interpolation. A mass resolution (M/ΔM) of c. 5400 was used to ensure adequate separation of Pb isotope peaks from nearby HfSi$^+$ species. Ion signals were detected using the axial ion-counting electron multiplier. All analyses were run in fully automated chain sequences. Data reduction assumes a power law relationship between Pb$^+$/U$^+$ and UO$_2^+$/$U^+$ ratios with an empirically derived slope in order to calculate actual Pb/U ratios based on those in the 91500 standard. U concentrations and Th/U ratio are also referenced to the 91500 standard. Common Pb corrections are made only when $^{204}$Pb counts statistically exceed average background and assume a $^{207}$Pb/$^{206}$Pb ratio of 0.83 (equivalent to present day Stacey and Kramers (1975) model terrestrial Pb). Decay constants follow the recommendations of Steiger and Jäger (1977). All age calculations were done in Isoplot 3.70 (Ludwig, 2008) and results are presented in Table 1.
4.2. Zircon Lu-Hf—isotope analyses

Lu-Hf analyses were carried out at the Advanced Analytical Centre at James Cook University in Townsville, Australia using a GeoLas 193-nm ArF laser and a Thermo-Scientific Neptune multi collector ICP-MS. Back scattered electron (BSE) images from a scanning electron microscope (SEM), transmitted and reflected light images were used to determine the optimum location of the spot on each zircon. Where possible, the Lu-Hf spots overlapped pits from the U-Pb analyses and spot sizes with a diameter of 31–58 μm were used. The interpreted crystallisation age of the individual sample was used in all Hf-isotope calculations. This age was also assumed for all undated (Lu-Hf isotope-) analysed grains of similar BSE character.

Each analysis began with a 30 second electronic baseline followed by an ablation period of 60 seconds involving 60 integration cycles of one second each. A laser pulse repetition rate of 4 Hz was used and the laser energy was held at ~6 J/cm² which equals an ablation rate of 0.06 μm per pulse for zircon. Helium carrier gas was used to transport the ablated particles from the sample chamber. It was combined with argon gas (flow rate ~0.8 l/min) and nitrogen (~0.005 l/min) further downstream before entering the argon plasma.

Masses 171 (Yb), 173 (Yb), 175 (Lu), 176 (Hf+Lu+Yb), 177 (Hf), 178 (Hf), 179 (Hf) and 180 (Hf+W+Ta) were measured simultaneously by Faradays detectors. Isobaric interference of 176Yb and 176Lu on 176Hf was calculated using the measured intensities of 171Yb and 175Lu along with known isotopic ratios of 176Yb/171Yb = 0.897145 (Segal et al. 2003) and 176Lu/175Lu = 0.02655 (Vervoort et al. 2004). Mass bias corrections were calculated using the exponential law. For calculations of βHf, measured intensities of 179Hf and 177Hf and a 179Hf/177Hf ratio of 0.7325 was used. βYb was calculated using measured intensities of 173Yb and 171Yb and a 176Yb/171Yb ratio of 1.130172 (Segal et al. 2003). Mass bias behaviour of Lu was assumed to be identical to Yb.

Three standards were used for quality control, FC-1, Mud tank zircon (Woodhead and Hergt 2005), and synthetic zircon (Fisher et al. 2011) and yielded 176Hf/177Hf of 0.282189 ± 0.00004 (2SD,
n=34), $^{176}\text{Hf} / ^{177}\text{Hf}$ of $0.282500 \pm 0.00003$ (2SD, n=55) and $^{176}\text{Hf} / ^{177}\text{Hf}$ of $0.282134 \pm 0.00003$ (2SD, n=24) respectively. These ratios are well within the range of solution mode data (Woodhead and Hergt 2005; Fisher et al. 2011) of FC-1= $^{176}\text{Hf} / ^{177}\text{Hf}$ of 0.282184 ± 16; Mud tank= $^{176}\text{Hf} / ^{177}\text{Hf}$ of 0.282507 ± 6; Fisher synthetic=0.282135 ± 7. In addition, the stable Hf isotope ratios, $^{178}\text{Hf} / ^{177}\text{Hf}$ and $^{180}\text{Hf} / ^{177}\text{Hf}$, were monitored since these should be constant within error throughout the measurements. Analysed $^{176}\text{Hf} / ^{177}\text{Hf}$ ratios of the unknown zircon grains were normalized based on comparison between the mean of analysed $^{176}\text{Hf} / ^{177}\text{Hf}$ ratios of Mud tank zircon and its reported $^{176}\text{Hf} / ^{177}\text{Hf}$ ratio of 0.282507 determined by solution analysis (Woodhead and Hergt 2005).

Calculations of $\varepsilon_{\text{Hf}}$ use $\lambda_{^{176}\text{Lu}} = 1.867 \times 10^{-11}$ yr$^{-1}$ (Scherer et al. 2001; Söderlund et al. 2004), $(^{176}\text{Lu} / ^{177}\text{Hf})_{\text{CHUR}} = 0.0336$ and $(^{176}\text{Hf} / ^{177}\text{Hf})_{\text{CHUR}} = 0.282785$ (Bouvier et al. 2008). Two stage model ages were calculated using new crust values of $^{176}\text{Hf} / ^{177}\text{Hf} = 0.28315$ and $^{176}\text{Lu} / ^{177}\text{Hf} = 0.03795$ (Dhuime et al. 2011) and by assuming a $^{176}\text{Lu} / ^{177}\text{Hf}$ of 0.0093 for the crustal source.

Results are presented in table 2. Secondary standard analyses are shown in supplementary figure A.1 and listed in supplementary table A.2.

5. Results

5.1. PK101 Amasaman biotite hornblende tonalite

Fourteen spots from oscillatory-zoned zircon cores were analysed. Two of these are concordant while remaining twelve spots define a discordia with intercepts at $2.126 \pm 0.012$ Ga and $0.500 \pm 0.057$ Ga respectively (MSWD=2.2; Fig. 11a). The upper intercept is interpreted to date the crystallisation age of this sample, while the lower intercept is in accordance with Pan-African Pb-loss in the response to the Dahomeyan orogen <10 km to the southeast. U and Th/U range between 267–628 ppm and 0.07–0.84 respectively. One analysis n3762-03 was discarded due to high common Pb ($^{206}\text{Pb} / ^{204}\text{Pb}=124$) and associated large error.
Nine Hf isotope analyses (of which #10 was discarded due to the laser penetrating the grain) yield $^{176}\text{Lu}/^{177}\text{Hf} < 0.0008$, $^{176}\text{Yb}/^{177}\text{Hf} < 0.03$ and $^{176}\text{Hf}/^{177}\text{Hf}$ from 0.281338 to 0.281416. The corresponding $\varepsilon\text{Hf}(2.126 \text{ Ga})$ values range between -4.2 and -1.3 (Fig. 11).

### 5.2. PK102 Nsawam biotite hornblende granite

In this sample, only oscillatory-zoned zircon core domains were analysed. A regression of all data points ($n=16$) yield a lower intercept of $0.523 \pm 0.096$ Ga, which points to Pan-African Pb-loss, and an upper intercept of $2.174 \pm 0.006$ Ga (MSWD=2.5; Fig. 11b), which is interpreted as the igneous crystallisation age of this sample. U concentrations range between 150–545 (150–425 ppm for data points used for concordia calculation) and Th/U range between 0.30–0.59 (0.38–0.59 for data points used for concordia calculation).

Nineteen Hf isotope analyses from eighteen grains yield $^{176}\text{Lu}/^{177}\text{Hf} < 0.0023$, $^{176}\text{Yb}/^{177}\text{Hf} < 0.07$ and $^{176}\text{Hf}/^{177}\text{Hf}$ range from 0.281454 to 0.281590. $\varepsilon\text{Hf}(2.174 \text{ Ga})$ ranges between +0.7 and +5.2 (Fig. 11).

### 5.3. PK103 Gomoa Fetteh hornblende biotite granite

Sixteen analyses of oscillatory zoned core domains were analysed. One slightly discordant spot ($n=15$) with a $^{207}\text{Pb}/^{206}\text{Pb}$-date of $2.460 \pm 0.015$ Ga is of xenocrystic origin. Remaining spots define a discordia with intercepts at $2.139 \pm 0.005$ Ga and $0.431 \pm 0.110$ Ga (MSWD=1.5; Fig. 11). The $2.139 \pm 0.005$ Ga intercept is interpreted as the igneous crystallisation age of this sample. U concentrations range between 38–382 ppm and Th/U range between 0.31–1.35, with no correlation with discordance.

Fourteen Hf isotope analyses of magmatic domains (two were discarded) yield $^{176}\text{Lu}/^{177}\text{Hf} < 0.0016$ and $^{176}\text{Yb}/^{177}\text{Hf} < 0.05$ and $^{176}\text{Hf}/^{177}\text{Hf}$ range from 0.281340 to 0.281516. $\varepsilon\text{Hf}(2.139 \text{ Ga})$ ranges between -3.8 and +1.7 (Fig. 11).

### 5.4. PK105 West Accra biotite hornblende granodiorite
Fifteen spots are discordant beyond the $2\sigma$-level and might represent a combination of Pan-African and recent Pb-loss (Fig. 11). In order to avoid the Pan-African overprint, concordant data with $(^{206}\text{Pb} / ^{204}\text{Pb} > 10000)$ were used to calculate a weighted average $^{207}\text{Pb} / ^{206}\text{Pb}$-date, which yielded $2.229 \pm 0.004$ Ga (MSWD=0.7; n=12/13; Fig. 11). We interpret this date as the best estimate of the igneous crystallisation age. U concentrations range between 101–638 ppm with a negative correlation between U concentration and $^{207}\text{Pb} / ^{206}\text{Pb}$-date. All analyses used to calculate the concordia age have <250 ppm U. Th/U for all analyses range between 0.09–0.64.

Twenty-eight Hf isotope analyses from 26 different grains yield $^{176}\text{Lu} / ^{177}\text{Hf} < 3.1 \times 10^{-3}$ and $^{176}\text{Yb} / ^{177}\text{Hf} < 0.08$ and $^{176}\text{Hf} / ^{177}\text{Hf}$ range from 0.281499 to 0.281645. Corresponding $\varepsilon_{\text{Hf}}(2.229 \text{Ga})$ values range between +2.0 and +6.3 (Fig. 11). Two analyses (-09, -10) were discarded, both due to short analysis time.

5.5. ASGH003A Cape coast two-mica granodiorite

Twenty-two analyses from different domains form a loosely defined discordia with intercepts at $2.097 \pm 0.041$ Ga and $0.408 \pm 0.030$ Ga respectively (MSWD=9.6). Three analyses are concordant, all from BSE bright oscillatory zoned domains, and yield a $^{207}\text{Pb} / ^{206}\text{Pb}$-date of $2.125 \pm 0.018$ Ga (MSWD=1.8; Fig. 11). This is interpreted as the igneous crystallisation age and is older than the 2.090-2.095 Ga age bracket given by 2.090 ±0.002 Ga monazite and slightly discordant 2.095 ±0.0034 Ga zircon, where Pan-African Pb-loss was not accounted for (Davies et al., 1994). U concentrations range between 579–5726 ppm (588–855 ppm for concordant data-points) and Th/U range between 0.00–0.10 (0.55–0.83 for concordant data-points). There is a negative correlation between U concentration and $^{207}\text{Pb} / ^{206}\text{Pb}$-dates.

Twenty-three Hf isotope analyses in 21 grains of which two where discarded due to the laser drilling thorough the grains (n3682-Hf-06, -13) and one (n3682-1b) due to heterogeneous $^{176}\text{Hf} / ^{177}\text{Hf}$ signal, yield $^{176}\text{Lu} / ^{177}\text{Hf} < 1.6 \times 10^{-3}$ and $^{176}\text{Yb} / ^{177}\text{Hf} < 0.05$ and $^{176}\text{Hf} / ^{177}\text{Hf}$ ranges from
0.281447 to 0.281620. The $\varepsilon_{\text{Hf}}(2.125 \text{ Ga})$ values range between -0.1 and +5.4 with a majority of the data (n=18) clustering between +2.1 and +4.3 (Fig. 11).

5.6. ASGH007A Dixcove hornblende-granite

Twelve analyses of oscillatory-zoned core domains yield data that are between 2.5–71.1% discordant beyond the 2σ-level. There is a clear trend with increased U concentration and discordance in domains with strong zonation. Discarding the three most discordant and U-rich analyses, all from strongly zoned domains, a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$-date of 2.173 ±0.012 Ga (MSWD=1.4, probability=0.2; Fig. 11) is obtained. Our date is in excellent agreement with the 2.172 ±0.002 Ga date obtained by Hirdes et al., (1992), and we interpret this as the igneous crystallisation age of the granite. U concentrations range between 71–1291 (71–118 ppm for concordant data points) and Th/U range between 0.03–0.51 (0.31–0.51 for concordant data points).

Nine Hf isotope analyses yield $^{176}\text{Lu}/^{177}\text{Hf} < 0.8 \times 10^{-3}$ and $^{176}\text{Yb}/^{177}\text{Hf} < 0.03$ and $^{176}\text{Hf}/^{177}\text{Hf}$ range from 0.2814479 to 0.281573. $\varepsilon_{\text{Hf}}(2.173 \text{ Ga})$ values range between +1.3 and +5.2 (Fig. 11).

5.7. ASGH022A Sunyani basin two-mica granite

Eleven oscillatory zoned core domains were analysed, of which all but two are concordant within error. The data define a discordia with intercepts at 0.152 ±0.260 Ga and 2.093 ±0.002 Ga (MSWD =0.9; Fig. 11) which is compatible with a recent Pb-loss model. The weighted average $^{207}\text{Pb}/^{206}\text{Pb}$-date of all spots yield a 2.093 ±0.002 Ga (MSWD=0.9; n=11/11; Fig. 11), and is interpreted as the igneous crystallisation age of this sample. U concentrations and Th/U range between 101–1273 and 0.07–0.67 respectively.

Hf isotope analyses (n=9) yield $^{176}\text{Lu}/^{177}\text{Hf} < 0.3 \times 10^{-3}$ and $^{176}\text{Yb}/^{177}\text{Hf} < 0.01$ and $^{176}\text{Hf}/^{177}\text{Hf}$ range from 0.281551 to 0.281587. Corresponding $\varepsilon_{\text{Hf}}(2.093 \text{ Ga})$ values range between +3.4 and +4.9 (Fig. 11).
5.8. ASGH022C Sunyani basin pegmatite

Seven analyses of BSE-bright oscillatory zoned core domains yield a discordia with only one intercept at 2.082 ±0.010 Ga (MSWD=1.1; Fig. 11). The weighted average $^{207}\text{Pb}/^{206}\text{Pb}$-date of all spots yield a 2.092 ±0.004 Ga (MSWD=1.2; $n=7/7$; Fig. 11), which is interpreted as the crystallisation age of this sample. U concentrations and Th/U range between 239–447 and 0.25–0.43 respectively.

Seven Hf isotope analyses yield $^{176}\text{Lu}/^{177}\text{Hf} <0.3\times10^{-3}$ and $^{176}\text{Yb}/^{177}\text{Hf} <0.01$ and $^{176}\text{Hf}/^{177}\text{Hf}$ range from 0.281547 to 0.281606. The $\varepsilon_{\text{Hf}}(2.092\text{ Ga})$ values range between $+3.2$ and $+5.5$ (Fig. 11).

6. Discussion

6.1. Juvenile granitic crust within the Birimian terrane

At the present day, the West African Craton is cut by, and juxtaposed with, juvenile Pan-African (Dahomeyan) crust in the southeast (e.g. Affaton et al., 1991). The paleo-extent of this Craton is unknown. However, as documented here, granite ages extend to >2.2 Ga towards its eastern margin, which are among the oldest within the Eburnean orogeny, and predate most mafic volcanic suites elsewhere in the Birimian terrane. The mafic volcanism has been ascribed to the arrival of a mantle plume (Abouchami et al., 1989) as well as subduction related volcanism (Sylvester and Attoh, 1992). Irrespective of tectonic model, the mafic magmatism is considered to represent juvenile crust generation between 2.15 to 2.2 Ga. To this end, it is notable that the >2.2 Ga granite magmatism that is documented here through sample PK105 has $\varepsilon_{\text{Hf}}(2.229\text{ Ga}) = +2.0-+6.3$, in line with estimates for the sub-lithospheric Birimian mantle from mafic volcanic rocks (Blichert-Toft et al., 1999), and implying derivation from juvenile crust.

More recently, it has been argued that Eoeburnean (c. 2.35–2.15 Ga) rocks have equivalents in various parts of the West African Craton and in the Brazilian São Luis Craton (deKock et al., 2011). These rocks are thought to correspond to a long-lasting period of juvenile crust formation (deKock...
et al., 2011). This is seen in the Eoeburnean Hf isotopic record where all combined zircon U-Pb–Hf
data yield juvenile supra-chondritic $\epsilon_{\text{Hf}}$ values (Fig. 12). Eoeburnean rocks crop out in an area
extending from southwestern Ivory Coast and Liberia to Burkina Faso and Ghana, with a few
occurrences of Eoeburnean rocks reported from eastern Guinea (Lahondère et al. 2002) and
southern Mali (McFarlane et al., 2011). Based on inherited 2.312 ±0.02 Ga zircon and literature
Sm-Nd model ages, Gasquet et al. (2003) proposed an early phase of crustal growth within the
Baoulé Mossi around 2.3 Ga. Early onset of the Birimian event has also been argued for by
Feybesse et al. (2006) based on ~2.35 Ga rocks within the Brazilian Boromea belt.
This early stage of evolved magmatic activity in the Birimian event contradicts the global 2.45–2.20
Ga magmatic quiescence proposed by e.g. Condie et al. (2009) but is in line with the more recent
views of Partin et al. (2014) who argue for uninterrupted Palaeoproterozoic plate tectonics.
Feybesse et al. (2006) suggests that juvenile crust formed during the Eoeburnean phase was
thickened through accretion between 2.16–2.15 Ga, coeval with the emplacement of large volumes
of monzonitic plutonic complexes found both in southern Ghana and in the São Luis Craton.
Between 2.15–2.10 Ga several basins (e.g. Sunyani, Kumasi-Afema and Comoé basins) formed
during an extensional tectonic regime (Feybesse et al., 2006). The initial part of this extensional
phase is coeval with a narrow span in crystallisation ages between 2.14–2.13 Ga that drop to sub-
chondritic $\epsilon_{\text{Hf}}$ values (Fig. 12). A similar drop is observed in detrital zircon data (Kristinsdóttir,
2013). This suggests that the reworking of Archaean crust within the Birimian terrane is limited to
this time-slice, and that it was preceded and succeeded by juvenile continental crust formation with
minimal or no contamination by older crust. This is in line with the detrital zircon record, which is
dominated by 2.15–2.06 Ga crystallisation ages and juvenile isotopic signatures (Kristinsdóttir,
2013; Izuka et al., 2013). Further work to explore the amount of reworked crust elsewhere in the
West African Craton is, however, required.

6.2. Reworking of Archaean material within the Birimian terrane
Our new zircon Lu-Hf data for c. 2.14-2.13 Ga granites from the Suhum basin display predominantly negative $\varepsilon$Hf, indicating significant involvement of older, tentatively Archaean, reworked crust (Fig. 12). This result corroborates whole rock Nd isotope data from the Winneba pluton in the Kibi-Winneba belt that yield a model age of c. 2.6 Ga (Leube et al., 1990; Taylor et al., 1992). Our new data extends the area where an Archaean signature is identified to include the Suhum basin southeast of the Kibi-Winneba belt (Fig. 12). Recalculating the Nd isotope data of Taylor et al. (1990) to $\varepsilon$Hf using equation: $\varepsilon$Hf = $1.55 \times \varepsilon$Nd + 1.21, as suggested by Vervoort et al. (2011) the Winneba pluton yields $\varepsilon$Hf = -7.2 (Fig. 12). This is even lower than the zircon data obtained here, but independent Hf isotope data or further work is required to test the validity of this correlation. We calculate two stage model ages using the measured $^{176}$Lu/$^{177}$Hf and the age of the zircon for the first stage, and an assumed $^{176}$Lu/$^{177}$Hf value of 0.0093 and the new crust curve of Dhuime et al. (2011) as a depleted mantle reference for the second stage. Considering the $\varepsilon$Hf$_{2.150\text{Ga}}$ $\approx$ 6 ± 2 estimate of the Birimian mantle provided by Blichert-Toft et al. (1999), a moderately depleted mantle evolution as suggested by Dhuime et al. (2011) or Iizuka et al. (2013) seems justified. The most enriched samples (PK101 and PK103) yield 2.4–2.7 Ga model ages (Table 2). In addition to Lu-Hf based model ages, a xenocrystic zircon with a $^{207}$Pb/$^{206}$Pb-date of 2.460±0.015 Ga was found in sample PK103 (Fig. 1b, Table 1), providing additional evidence for the reworking of older crust. Irrespective of mantle model, a majority of the analysed grains from southern Ghana require reworking of an ancient component to explain their Hf isotope ratios. This suggests a more substantial contribution of reworked Archaean crust to the southern parts of the Birimian terrane in Ghana than previously known. Detrital zircon grains from the Cadomian Orogen in central west Europe include a 1.8–2.2 Ga component that is interpreted to have derived from the West African Craton (Linnemann et al., 2014). The model ages of this population imply reworking of a 2.5–3.4 Ga basement, using a MORB-mantle depletion model. Furthermore, detrital zircon from the Anti-Atlas belt in southern
Morocco have an Archaean component with Hf model ages varying between 2.3 and 3.3 Ga (Abati et al., 2012). The origin of these grains is unknown, but the agreement between the Anti-Atlas zircon model ages and the least radiogenic data from southern Ghana opens for the possibility of a Birimian source to these zircon grains. However, the inference about the antiquity of the West African Craton by Linnemann et al. (2014) is only partly conceivable when compared with our results, where significant reworking of ancient crust appears to be limited to a period between 2.14 to 2.13 Ga. Their conclusion is in stark contrast with the generally juvenile nature of Birimian rocks, which is supported by our data as well as having been noted in previous studies (e.g. Abouchami et al., 1990; Liégeois et al., 1991; Boher et al., 1992; Ama-Salah et al., 1996; Hirdes et al., 1996; Doumbia et al., 1998; Gasquet et al., 2003; Pawlig et al., 2006; Klein et al., 2008; Tapsoba et al., 2013). Further study is required to establish the degree as well as the spatial and temporal distribution of reworking of Archaean crust across the West African Craton.

6.3. On the scarcity of xenocrystic zircon

The small number of pre-Eburnean xenocrystic zircon found in this study (n = 1) and within the Birimian terrane of the West African Craton as a whole (n ≈ 40; c.f. De Kock et al., 2011) is curious given our Hf isotope evidence for reworking of ancient crust (Fig. 12). This might be explained by a zircon poor or absent protolith, reflect biased sampling or physiochemical properties of magmas that caused resorption of inherited zircon.

The phenomenon with a few zircon xenocrysts in rocks that have enriched isotope signatures, indicating reworked older crust, is not unique to the Birimian terrane. Similar observations are made both in regional and global datasets. For example, Eoarchaean to Neoarchaean basement rocks in southern West Greenland with variably enriched zircon Hf isotope signatures that were interpreted to have crystallised from reworked older continental crust lack or have few xenocrystic zircon (Hiess et al., 2011; Næraa et al., 2012; 2014). In the case of the 2.55 Ga Qorqût granite in southern
West Greenland, Næraa et al., (2014) argued that the source rock was Eoarchaean mafic crust, which likely would supply few xenocrystic zircon grains to the magma. Palaeo- to Mesoproterozoic intrusions in southern Fennoscandia that intrude and rework metasedimentary basins have few xenocrystic zircon grains (Petersson et al., 2015a; 2015b). In these two studies, the scarcity of xenocrystic zircon might in part be due to sampling bias as euhedral simple magmatic zircon was targeted (Petersson et al., 2015b), or the alkaline nature of some magmas might have dissolved zircon to a higher extent (Petersson et al., 2015a). In contrast to these studies, a large number of xenocrystic zircon was retrieved from rocks crystallised from initially zircon-undersaturated magmas within the Phanerozoic Lachland Orogen (Kemp et al., 2005).

On a global scale, there is a similar enigmatic discrepancy between the small amount of pre-3.0 Ga zircon (ca. 10%) and the large inferred mass fraction of continental crust (50 – 70% of the present mass; Belousova et al., 2010; Dhuime et al., 2012). To what extent the scarcity of xenocrystic zircon within the Birimian terrane represent sampling bias, source characteristics or zircon dissolution due to physiochemical magma properties remains unclear.

6.4. Birimian isotopic signatures in a tectonic context

The Birimian crust is a commonly cited example (e.g. Arndt, 2013) of plume-related crustal growth, where the mafic volcanism has been proposed to represent the first stage of the crustal evolution (Abouchami et al. 1990; Vidal et al. 1996; Doumbia et al. 1998; Lompo 2009, 2010; Vidal et al. 2009). Boher et al. (1992) propose a model where the Birimian crust initially formed a plume-related oceanic plateau around which subduction zones subsequently reworked the oceanic plateau before it was accreted to the Archaean nucleus of the Man Shield. The main arguments for this interpretation include – the common occurrence of pillow lavas and the absence of rocks with
affinities of the continental crust, the juvenile isotopic character of the Birimian terrane and the
geochemical signatures of the Birimian mafic supracrustal rocks.

In contrast, other workers have argued for a subduction setting for basaltic and andesitic rocks
within the Birimian crust (e.g. Sylvester and Attoh, 1992; Evans et al., 1996; Ama Salah et al.,
1996; Baratoux et al, 2011). The juvenile character of the Birimian terrane is the single uniting
interpretation, which is based on the scarcity of xenocrystic zircon and Sr, Nd and Hf isotopic
compositions that indicate purely juvenile crustal growth.

If a mantle plume model is based on characteristics of comparatively well-established Phanerozoic
analogs such as the Deccan–Reunion or the Parana–Etendeka–Tristan da Cunha, the main eruptive
stage of flood basalt volcanism should last for c. 1 Myr (Shoene et al., 2015; Thiede and
Vasconcelos, 2010). In contrast, the Birmian is characterised by at least two >5 Ma pulses of
basaltic magmatism, which are separated by ~35 Myr (Fig. 12, Abouchami et al., 1990; Sylvester
and Attoh, 1992; Vidal and Alric, 1993; Dampare et al., 2008; Baratoux et al., 2011). Furthermore,
as shown here, emplacement of evolved granitic rocks (PK105, West Accra biotite hornblende
granodiorite) predates the mafic-ultramafic volcanism in the Birimian terrane, which contradicts the
hypothesis of a plume-initiated growth cycle (Fig. 12). Our new zircon isotope data also negate the
hypothesis presented by Boher et al. (1992), suggesting assimilation of older crust during anatexis,
and crust generation in close proximity to existing continental crust.

The available literature data for Birimian rocks have somewhat contrasting geochemical signatures,
where mafic rocks are akin to ocean floor basalt, while the felsic rocks are dominated by magnesian
granitic rocks with arc-like trace element signatures. To this end, it is worth noting that
discriminating tectonic setting solely based on geochemical signatures has shortcomings unless
these signatures are uniquely linked to physical processes (e.g. Hawkesworth and Scherstén, 2007).
Nevertheless, taking chronological and geochemical data into account, the ocean plateau model
proposed by Boher et al. (1992) seems untenable as the mafic magmatism is preceded and
interleaved by calc-alkaline, magnesian felsic magmatism. By modern analogy, the mafic plateau-
building stage should rather have been represented by a short period with a large volume eruptive phase that preceded felsic magmatism. The alternative arc accretion model (Sylvester and Attoh, 1992; Feybesse and Milési, 1994; Ama-Salah, et al. 1996; Pouclet et al., 2006; Baratoux et al., 2011; de Kock et al., 2012) is more in line with available data, where some of the mafic magmatic stages might represent extensional periods of back-arc magmatism.

6.5. Alternating tectonics during crustal growth of the Birimian terrane

The temporal $\varepsilon$Hf-trends can be put into a plate tectonic framework with eastward subduction in a predominantly retreating arc system (Fig. 13). It is envisaged that juvenile island arc magmatism dominates between ~2.35–2.20 Ga (Fig. 13a). During this time period the West Accra granodiorite, PK105 crystallised (Fig. 12). Accretion of this island arc system to an assumed Archaean crust between ~2.18–2.13 Ga led to the crystallisation of PK102, ASGH007A, PK103 and PK101 (Fig. 13b). The ~2.18–2.13 Ga magmatism incorporates crust from an assumed Archaean terrane to the east as reflected by the subchondritic Hf isotope signatures seen in figure 12. The 2.17 Ga Nsawam biotite hornblende granite (PK102) has slightly less depleted $\varepsilon$Hf values than the contemporaneous Dixcove tonalite (ASGH007A) to the west (Figs. 12 and 13b). These differences might reflect trench-ward magmatism without reworked Archaean crust in the Dixcove tonalite while retro-arc magmatism to the east might have involved reworked Archaean crust. The pronounced Archaean influence between 2.141–2.126 Ga, as seen in the Gomoa Fetteh hornblende biotite granite (PK103) and the Amasaman biotite hornblende tonalite (PK101) samples (Fig. 12), coincides with the peak in Birimian crystallisation ages and argues for a continental setting during emplacement of these rocks. At ~2.13 Ga the main Eburnean orogeny began (Leube et al., 1990; Eisenlohr and Hirdes, 1992; Hirdes and Davis, 1998; Feybesse et al., 2006), and between 2.15–2.10 Ga several basins formed during an extensional phase (Feybesse et al., 2006), potentially explaining the abrupt return to supra-chondritic Hf-isotope signatures (Fig. 13c–d). This stage might have been associated with slab retreat and trench-ward magmatic migration from a thickened retro-arc into the thinned
extension zone where mantle derived magmas mix with juvenile continental crust generating melts with juvenile Hf isotope signatures (Kemp et al. 2009).

Alternatively, crustal thickening during the closure of oceanic back-arc can bury metasedimentary rocks derived from the Craton that melt during a subsequent extensional phase, giving rise to distinct but brief (<50 Myr) excursions toward negative Hf isotopic signatures (Bahlburg et al., 2009; Kemp et al. 2009; Mišković and Schaltagger, 2009; Collins et al., 2011). Such a scenario would, however, require the Archaean source to derive from sedimentary rocks, and all detrital zircon grains with sub-chondritic Hf isotope signatures reported by Kristinsdóttir, (2013) have more or less mantle oxygen signatures, suggesting that the Archaean crust never interacted with the hydrosphere. It is also noteworthy that samples with sub-chondritic Hf isotope signatures in this study are hornblende-bearing (metaluminous) granites, arguing against a S-type origin.

Although intrusions that are younger than 2.13 Ga are relatively radiogenic for Hf (Fig. 12), they likely contain a component of ~2.3–2.2 Ga juvenile crust, as they host abundant metasediment xenoliths and are two-mica granites with a strong peraluminous signature.

7. Conclusions

The contribution from Archaean crust to the Birimian terrane is greater than previously known and comprises not only the Winneba pluton but also larger parts of the Kibi-Winneba belt as well as rocks intruding the Suhum basin. Reworking of Archaean crust was active during a short time period between ~2.14–2.13 Ga, where preceding and subsequent magmatism has relatively juvenile character.

The 2.23 Ga age of the West Accra granodiorite (PK105) requires the emplacement of felsic crust during the Eoeburnean and pre-dates suggested plume related rocks of Abuchami et al. (1990) and Boher et al. (1992) contradicting a suggested plume-initiated crustal growth stage.
An eastward, mainly retreating arc system with a shorter pulse of accretion between ~2.18–2.13 Ga and a rapid return to slab retreat explains trends seen in the combined zircon U-Pb and Lu-Hf isotope data and the geographical propagation of Archaean contribution to Birimian rocks.


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Figure Captions

**Fig. 1a.** Simplified tectonic map of the West African Craton and adjacent Pan-African-Hercynian fold and thrust belts. Mesoproterozoic to recent sedimentary rocks are not depicted. The map has been compiled from the following sources; Man-Leo shield, Kedougou-Kénia, Kayes (Egal et al. 2002; Baratoux et al. 2011), Reguibat shield (Peucat et al. 2005; Schofield et al. 2012), Pan-African belts (Persits et al. 2002; Baratoux et al. 2011) and Hercynian belt (Abouchami et al. 1990; Schofield et al. 2012). WAC boundaries after Ennih and Liégeois (2008). Redrawn after Grenholm (2014).


**Fig. 1c.** Geological map of Ghana showing sample locations, basins, belts and main rock units. Initial version of the map was compiled by Watts, Griffit and McQuat Ltd, Lakewood Colorado, USA.

**Fig. 2.** Small samples aliquot (left) showing macroscopic features. Plane polarised thin section view (ppl) of a representative area (middle). Cross polarised thin section view (xpl) of the same area as for the plane polarised view (right).
Fig. 3. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 4. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 5. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 6. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 7. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 8. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.
Fig. 9. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 10. BSE (Back-Scattered-Electrone) image of representative zircon grains. Ellipses indicate spot locations, small thin: U-Pb and large thick: Lu-Hf. Numbers inside U-Pb ellipses refer to analytical ID in U–Pb and Lu–Hf data tables. Dashed ellipses and results in italic denote discarded analyses.

Fig. 11. Tera-Wasserburg concordia diagrams showing SIMS (Secondary-Ion-Mass-Spectrometry) zircon spot data for all samples (±2 error ellipses) and obtained ages. All ages are shown with 2 errors. Red ellipses denote discarded analyses not used in age calculation. Dashed lines denote discordia lines.

Fig. 12. εHf versus crystallisation ages (in Ma). εHf has been calculated using current CHUR values of $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ from Bouvier et al. (2008). $^{176}\text{Lu}$ decay constants of Söderlund et al. (2004) and Scherer et al. (2001) were used in all calculations. Ages represent interpreted igneous crystallisation ages for individual samples. εHf-value of the Winneba pluton corresponds to the recalculated Nd-isotope data of Taylor et al. (1990), including age error bars. Vertical grey lines represent timing of reported mafic volcanism in the Baoulé Mossi domain (Abouchami et al., 1990; Sylvester and Attoh, 1992; Vidal and Alric, 1993; Dampare et al., 2008; Baratoux et al., 2011).

Fig. 13. Theoretical evolutionary model proposed for the arc system generating the Birimian terrane in Ghana. A. Retreating eastward subduction generating juvenile island arc magmatism outboard
the Western Archaean crust. B. Switch to an advancing arc system with accretion of the juvenile island arcs onto the eastern Archaean crust. ~2.18–2.13 Ga magmatism incorporates crust from the Archaean nucleus to the east as reflected in subchondritic Hf-isotope signatures. C. Slab retreat migrates igneous activity trench-ward from the thickened back arc into the thinned extension zone where mantle derived magmas mix with juvenile continental crust. D. Continued extensional tectonic regime and simultaneous amalgamation of the Birimian crust to the western Archaean Man-Shield.

Supplementary figure A.1. Mean values of standard runs during Hf-isotope analyses presented in $^{176}\text{Hf}/^{177}\text{Hf}$. Data quality was controlled using standards Mud Tank, FC-1 (Woodhead and Hergt 2005) and synthetic zircon (Fisher et al. 2011).
<table>
<thead>
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<th>Sample</th>
<th>Pb 206/207</th>
<th>Pb 207/206</th>
<th>Pb 208/206</th>
<th>Th/232U</th>
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<th>Age (Ma)</th>
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Table 1. U-Pb ages and ε (Pb) values for selected samples.
where the letter b is added to spot name it indicates a second spot in an already analysed grain.

Sample

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**Age discordance at closest approach of error ellipse to concordia (2σ)**
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<th>Error (Ma)</th>
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<th>Error (Ma) 3</th>
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Note: All values are given in Ma (million years before present). Error bars represent 1-sigma uncertainties.
Figure 3
Figure 4

PK102

$^{207}\text{Pb} / ^{206}\text{Pb} = 2185 \pm 4 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 3.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2174 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 3.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2121 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 3.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2179 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 2.6$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2182 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 2.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2183 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 1.4$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2159 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 1.4/2.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2179 \pm 4 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 0.8$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2111 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 1.5$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2162 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 2.9$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2182 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 2.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2156 \pm 4 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 5.3$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2169 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 2.1$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2180 \pm 6 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 3.0$

$^{207}\text{Pb} / ^{206}\text{Pb} = 2158 \pm 5 \text{ Ma}$
$\varepsilon_{\text{Hf}} = 3.4$

100 µm
Figure 5

PK103

$^{207}\text{Pb}/^{206}\text{Pb} = 2152 \pm 6 \text{ Ma} \quad \varepsilon\text{Hf}=1.6$

$^{207}\text{Pb}/^{206}\text{Pb} = 2146 \pm 7 \text{ Ma} \quad \varepsilon\text{Hf}=-3.6$

$^{207}\text{Pb}/^{206}\text{Pb} = 2109 \pm 9 \text{ Ma} \quad \varepsilon\text{Hf}=-2.4$

$^{207}\text{Pb}/^{206}\text{Pb} = 2132 \pm 10 \text{ Ma}$

$^{207}\text{Pb}/^{206}\text{Pb} = 2118 \pm 16 \text{ Ma} \quad \varepsilon\text{Hf}=-2.6$

$^{207}\text{Pb}/^{206}\text{Pb} = 2163 \pm 21 \text{ Ma} \quad \varepsilon\text{Hf}=-2.2$

$^{207}\text{Pb}/^{206}\text{Pb} = 2140 \pm 6 \text{ Ma}$

$^{207}\text{Pb}/^{206}\text{Pb} = 2164 \pm 9 \text{ Ma} \quad \varepsilon\text{Hf}=-2.5$

$^{207}\text{Pb}/^{206}\text{Pb} = 2134 \pm 6 \text{ Ma}$

$^{207}\text{Pb}/^{206}\text{Pb} = 2115 \pm 11 \text{ Ma} \quad \varepsilon\text{Hf}=-3.2$

$^{207}\text{Pb}/^{206}\text{Pb} = 2125 \pm 8 \text{ Ma} \quad \varepsilon\text{Hf}=-2.5$

$^{207}\text{Pb}/^{206}\text{Pb} = 2131 \pm 8 \text{ Ma} \quad \varepsilon\text{Hf}=-2.6$

$^{207}\text{Pb}/^{206}\text{Pb} = 2119 \pm 9 \text{ Ma} \quad \varepsilon\text{Hf}=-4.5/-6.1$

$^{207}\text{Pb}/^{206}\text{Pb} = 2460 \pm 15 \text{ Ma} \quad \varepsilon\text{Hf}=11.0$
Figure 9

ASGH022A

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2094 \pm 4 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.9 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2084 \pm 13 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.5 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2095 \pm 5 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.3 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2088 \pm 4 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.3 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2094 \pm 5 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.7 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2101 \pm 9 \text{ Ma} \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2095 \pm 5 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 4.3 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2088 \pm 6 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 3.9 \]

\[ \frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 2094 \pm 4 \text{ Ma} \]
\[ \varepsilon\text{Hf} = 3.9 \]