

The geodynamic and tectonic evolution of the Palaeoproterozoic Sefwi Greenstone Belt, West African Craton (Ghana)

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Évolution géodynamique et tectonique de la ceinture de roches vertes Paléoprotérozoïque de Sefwi, Craton Ouest-Africain (Ghana)

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The geodynamic and tectonic evolution of the Palaeoproterozoic Sefwi Greenstone Belt, West African Craton

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Abstract

This thesis investigates the Palaeoproterozoic crust of the West African Craton in southwest Ghana, providing insight into its controversial geodynamic and tectonic evolution. Rocks of the study area comprise greenschist- to amphibolite facies, mafic to felsic volcanic and volcaniclastic rocks, high-grade paragneisses and low-grade volcano-sedimentary packages, all of which are extensively intruded by multiple generations of granitoids. New lithological, metamorphic and structural maps are constructed using integrated field mapping and interpretation of regional airborne geophysical datasets. This approach is used to constrain the deformation history of the sparsely exposed rocks of the NE- to NNE-striking Sefwi Greenstone Belt and the adjacent volcano-sedimentary domains deformed during the Eburnean Orogeny (2150–2070 Ma).

Combined geochemical and geochronological analysis of the magmatic suites of the Sefwi Greenstone Belt reveal calc-alkaline, volcanic arc affinities, as well as a striking similarity to Neoarchean TTGs that require diverse magma sources and petrogenetic processes. Rare inherited zircon cores from the Palaeoproterozoic magmatic suite yield ages of ca. 2250 to 2270 Ma, with granitoid emplacement ages ranging between ca. 2189 and 2081 Ma. Zircon Lu-Hf analysis reveals consistently positive $\varepsilon H f_{(t)}$ values and two-stage crustal model ages between 2650 and 2250 Ma, indicative of a radiogenic proto-crust and short crustal residence times. The magmatic evolution reveals the coeval generation of sodic, high-silica TTGs derived from partial melting of low-K mafic sources and dioritic magmas generated in a metasomatised, LILE-enriched mantle wedge at ca. 2155 Ma. Subsequent emplacement of high-K quartz monzonites at ca. 2136 Ma supports the interaction of mantle-derived magmas and remelting of existing TTGs. The final stage of magmatism is characterised by the emplacement of two-mica-granites and leucogranites along the NW margin of the Sefwi Greenstone Belt between ca. 2092 and 2081 Ma, interpreted as a terminal collisional event during the Eburnean Orogeny.

Eburnean metamorphism and deformation is characterised in the study area by initial high-pressure amphibolite facies metamorphism corresponding with low apparent geothermal gradients (HP-MT, \sim 15–17°C/km). D1 NNW-SSE shortening generated a ubiquitous bedding-parallel foliation (S₁) and \sim E-W striking thrust faults, resulting in the burial of supracrustal rocks and crustal thickening. In the high-grade terrane, subsequent amphibolite-granulite facies

metamorphism is associated with anatexis. In-situ SHRIMP U-Pb monazite ages at ca. 2073 Ma, hosted within, D_2 mineral assemblages, are interpreted as the initial timing of cooling and exhumation, significantly later than paroxysmal metamorphism in NW Ghana and central Ivory Coast (2150–2130 Ma). NNE-striking normal detachments and constrictional deformation structures formed during sinistral ENW-WSW transtension (D_2), during which segments of the middle- and lower crust were juxtaposed with low-grade domains. Subsequent E-W directed shortening (D_3) caused the dextral re-activation of NE-SW striking shear zones, associated with a localised greenschist facies metamorphic overprint.

We propose that the juvenile crust of southwest Ghana was generated in an intraoceanic arc setting, associated with diverse and intense subduction-related magmatism until subsequent terrane accretion and collision. The north-western margin of the Sefwi Greenstone Belt in interpreted as a suture between the separate arc terranes, diachronously accreted during the Eburnean Orogeny. The Palaeoproterozoic crust of the southern portion of the West African Craton represents a juvenile crustal growth event, recording the unique geodynamic and orogenic processes associated with nascent subduction-related plate tectonics in the early Earth.

Résumé

Cette thèse s'intéresse à un segment de croûte d'âge Paléoprotérozoïque du craton ouestafricain. Les roches de la zone d'étude comprennent des roches volcaniques et volcanoclastiques mafiques à felsiques, des paragneiss de haut grade métamorphique et des ensembles volcano-sédimentaires faiblement métamorphisés. De nouvelles cartes lithologiques, métamorphiques et structurales sont construites à l'aide d'une approche intégrée, couplant cartographie de terrain et interprétation des données géophysiques aéroportées à l'échelle régionale.

L'analyse des données géochimiques et géochronologiques des suites magmatiques de la ceinture de roches vertes de Sefwi révèle une affinité marquée avec le magmatisme calcoalcalin, produit des arcs volcaniques modernes et avec les TTGs d'âge Néoarchéen, impliquant une certaine diversité des sources et des processus pétrogénétiques. Des coeurs de zircons hérités, présents au sein de la suite magmatique livrent des âges autour de ca. 2250 à 2270 Ma. Leurs couronnes révèlent des âges de mise en place compris entre ca. 2189 et 2081 Ma. L'analyses Lu-Hf sur zircon livre des valeurs ɛHf positives et des âges modèles pour la croûtes situés entre 2650 et 2250 Ma. Ces valeurs indiquent l'existence d'une proto-croûte à tendance radiogènique et des temps de séjour limités pour ces magmas évoluant au sein de cette protocroûte. L'évolution des magmas montre qu'ils ont été générés de façon concomitante, vers 2155 Ma, certains dérivants de la fusion d'une source mafique faiblement enrichit en potassium et formant des magmas sodiques, riches en silice, de type TTGs, d'autres de composition plus dioritiques, générés à partir de la fusion du manteau métasomatisé et enrichit en LILE. La mise en place plus tard vers ca. 2136 Ma de monzonites, présentant de teneurs élevées en potassium, soutient l'hypothèse d'une interaction avec des magmas de refusion de TTG existants au sein de la croûte. Le dernier stade du magmatisme est caractérisé par la mise en place de granites à deux micas et de leucogranites, le long de la marge nord-ouest de la ceinture de roches vertes vers ca. 2092 et 2081 Ma, marquant le stade de la collision au sein de l'orogène Eburnéenne.

L'évènement tectono-métamorphique d'âge Eburnéen est caractérisé par un métamorphisme initialement de faciès amphibolite de haute pression, associé à un gradient géothermique assez froid (HP-MT, ~ 15-17 ° C / km). Le raccourcissement D1, orienté NNO-SSE, a généré une foliation pénétrative (S1), parallèle au litage des roches et des plans de chevauchement à tendance décrochant, orienté E-O. Cette tectonique précoce a provoquée l'enfouissement de roches supra-crustales (sédiments, roches volcaniques) et un épaississement

de la croûte. Cet évènement métamorphique précoce évolue dans le temps et l'espace vers le facies amphibolite-granulite et l'anatexie. Les données SHRIMP U-Pb in-situ sur monazite livré des âges autour de ca. 2073 Ma. Ces monazites sont présentes au sein de paragénèses métamorphiques (D2) soulignant la foliation S2. Ces âges sont interprétés comme marquant le début de l'exhumation et du refroidissement de la croûte inférieure. Des détachements normaux, orientés NNE-SSO et des structures constrictives se sont formés conjointement au sein d'un régime de déformation D2 globalement transtensif, à jeux sénestre. Un régime compressif plus tardif (D3) a ensuite causé une réactivation en mouvement dextre de ces structures cisaillantes orientées NE-SO avec une rétrogression en schistes verts.

Nous proposons que les segments de croute juvénile ont été générés en contexte d'arc intra-océanique, associé à un magmatisme intense et varié, issus des processus de subduction qui prendront fin lors des stades d'accrétion et de collision de ces segments d'autres terranes birimiens. La marge nord-ouest de la ceinture de roches vertes de Sefwi est interprétée comme une zone de suture entre des segments d'arc originellement séparés.

Declaration

This thesis contains no material which has been accepted for the award of any other degree or diploma at any university or equivalent institution and that, to the best of my knowledge and belief, this thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

This thesis contains one paper submitted for publication to a peer-reviewed journal (*Thesis including published works declaration; p. vii*), and two additional chapters (below table) to be submitted for publication in the near future. The ideas, development and writing up of all the papers in the thesis were the principal responsibility of myself, the candidate, working within the School of Earth, Atmosphere and Environment, Monash University under the supervision of Dr. Laurent Aillères and Prof. Peter Betts, and Géosciences Environnement Toulouse (GET), Université Toulouse III - Paul Sabatier under the supervision of Prof. Mark Jessell, Dr. Jérôme Ganne and Dr. Lenka Baratoux. The inclusion of co-authors reflects the fact that the work came from active collaboration between researchers and acknowledges input into team-based research. In the case of Chapter 4, my contribution to the work involved the following:

Thesis	Publication title	Status	Nature and extent (%)	Contributors/co-authors in
chapter			of students	order of
			contribution	contribution
	Transtension-related lower crust		90%; Fieldwork, sample	Sylvain Block
	exhumation in the late stages of		preparation, data	Laurent Aillères
3	the Palaeoproterozoic Eburnean		processing,	Peter Betts
	Orogeny, SW Ghana: Evidence	Not	interpretation and	Jérôme Ganne,
	for diachronous assembly of the	submitted	analysis, preparation of	Mark Jessell
	West African Craton		manuscript	Lenka Baratoux
	Petrological and		90%; Fieldwork, sample	Sylvain Block
	geochronological evidence of		preparation,	Jérôme Ganne
	collisional orogenesis and lower	Returned for	petrographic analysis,	Lenka Baratoux
4	crust exhumation during the	revision and	mineral chemistry, P-T	Laurent Aillères
	Palaeoproterozoic Eburnean	re-	calculations,	Peter Betts
	Orogeny, SW Ghana, West	submission	preparation of	Mark Jessell
	African Craton		manuscript	Richard Armstrong
			90%; Fieldwork, sample	Luis A. Parra-Avila
	Palaeoproterozoic juvenile crust		preparation, mineral	Laurent Aillères
	formation of SW Ghana in the		separation, petrographic	Robin Armit
	West African Craton: New	Not	and geochemical data	Nicolas Thébaud
5	insights from igneous	submitted	analysis, preparation of	Christopher Spencer, Jérôme
	geochemistry and U-Pb-Hf		manuscript	Ganne, Peter Betts
	zircon data			Mark Jessell, Lenka Baratoux

I have renumbered sections of submitted or published papers in order to generate a consistent presentation within the thesis.

Helm Mel-lare

Student signature:

Date: 5/12/2017

The undersigned hereby certify that the above declaration correctly reflects the nature and extent of the student and co-authors' contributions to this work. In instances where I am not the responsible author I have consulted with the responsible author to agree on the respective contributions of the authors.

Main Supervisor signature:

Date: 5/12/2017

Résumé analytique (Executive summary)

Objectifs de la thèse

Les roches d'âge Paléoprotérozoïques du craton ouest-africain, datés autour de 2300-2070 Ma, sont actuellement sous-explorées et peu représentées dans la littérature mondiale qui s'intéresse à la géodynamique protérozoïque. Ce craton expose pourtant un ensemble unique des roches crustales formées après la transition archéen-protérozoïque et qui n'ont subi que très peu de modifications après la stabilisation de ce dernier. Ce craton offre ainsi une fenêtre exceptionnelle sur le cadre géodynamique et les processus orogéniques responsables de la formation et la conservation de la croûte juvénile et des processus orogéniques au cours d'une période énigmatique de l'histoire de la Terre. Dans le cadre de l'Initiative pour l'Exploration en Afrique de l'Ouest (WAXI), les chercheurs d'universités d'Afrique de l'Ouest, d'Australie, de France et du Canada se sont associés avec les acteurs industriels, des sponsors et les bureaux nationaux de recherche en géologie, au sein d'un programme intégré qui vise à améliorer nos connaissances sur la géologie et le potentiel d'exploration du craton ouest-africain. Ce projet représente l'un des derniers projets de doctorat à l'échelle régionale du programme WAXI au Ghana, ce qui représente une opportunité pour relier la technique de recherche récente et les résultats du sud et du nord-ouest du Ghana. La ceinture de roches vertes de Sefwi, d'âge Paléoprotérozoïque, située au sud-ouest du Ghana, constitue la zone ciblée par cette thèse. Les principales questions que cette thèse vise à aborder peuvent être listées comme suit:

1. Quelle a été l'évolution structurale et métamorphique de la ceinture de roches vertes de Sefwi?

2. Quelles ont été les caractéristiques de la déformation et du métamorphisme au Paléoprotérozoïque ? Quelles ont été les forces motrices à l'origine de cet évènement tectono-métamorphique ? En outre, qu'est-ce que cela nous apprend sur le style tectonique de l'orogenèse éburnéenne et quelles en sont les implications pour la compréhension de cette orogenèse Paléoprotérozoïque?

3. Quels ont été les paramètres géodynamiques qui ont permis la formation et la préservation de grands volumes de croûte juvénile au sein du craton ouest-africain, suite à une période hypothétique de quiescence tectonique globale? Est-ce que les processus géodynamiques responsables de la formation du craton ouest-africain ont des similitudes avec les processus archéens ou ont-ils été formés dans un cadre tectonique

différent, de type Tectonique des Plaques? Et en quoi ces processus de Tectoniques de Plaques diffèrent-ils de ceux d'aujourd'hui?

L'objectif principal de cette thèse est d'explorer le cadre géodynamique de la formation de la croûte juvénile ainsi que les processus orogéniques préservés au sein du craton ouest-africain pour avoir un aperçu de l'évolution du régime de la Tectonique des Plaques au début de l'histoire de la Terre.

Compte tenu de ces objectifs ambitieux et de la taille de l'objet d'étude, une approche multi-échelles et multi-disciplinaires a été privilégié, combinant données de terrain, cartographie régionale, analyse structurale et modélisation métamorphique, couplé à des datations et des analyses géochimies et isotopiques. Les grandes lignes de la thèse adressent différents aspects au sein de trois principaux objectifs, incluant le contexte au sein de la littérature existante et les méthodes appliquées afin d'explorer différents aspects de chaque question.

Les directeurs de thèse pour ce projet étaient le Dr. Laurent Aillères (School of Earth, Atmosphere and Environment, Monash University), le Prof. Mark Jessell (Centre for Exploration Targeting, University of Western Australia/ Géosciences Environnement Toulouse/ L'Institut de Recherche pour le développement), le Prof. Peter Betts (School of Earth, Atmosphere and Environment, Monash University), le Dr. Lenka Baratoux (GET/IRD) and le Dr. (HDR) Jérôme Ganne (GET/IRD). De proches collaborations ont pu être établie avec Sylvain Block (GET), Nicholas Thébaud (CET, UWA), Luis Parra-Avila (CET/UWA), Robin Armit (School of Earth, Atmosphere and Environment, Monash University) et Christopher Spencer (Earth Dynamics Research Group, Curtin University).

Ce projet a été financé par AMIRA International et les commanditaires de l'industrie, y compris AusAid et le projet ARC Linkage LP110100667, qui a bien voulu soutenir le projet WAXI (P934A). Une part importante de la recherche a été financée par le School of Earth Atmosphere and Environment, Monash Research Initiative Fund – à travers le Dr Laurent Ailleres et des contributions du Dr Robin Armit et du professeur Peter Betts, ainsi que par l'Institut de Recherche pour le Développement (IRD), à travers les fonds de recherche du Dr Jérôme Ganne.

Chapitre III Résumé

Ce chapitre se concentre sur les lithologies et les séquences de déformation préservées dans la ceinture de roches vertes du Sefwi et dans les domaines tectono-métamorphiques adjacents au sud-ouest du Ghana. Nous présentons de nouvelles cartes litho-structurales et métamorphiques de la zone d'étude, qui sont le produit de l'intégration de la cartographie sur le terrain, des données pétrophysiques obtenues sur des lithologies prélevées localement et des données géophysiques aéroportées obtenues à l'échelle régionale. L'étude structurale multi-échelle et l'intégration de plusieurs ensembles de données géologiques sur une région où les affleurements géologiques se font rares, permet de mieux appréhender l'histoire tectonique de ce craton. Nous considérons ici le style et l'histoire de la déformation dans le contexte des modèles tectoniques existants pour le craton ouest-africain et des modèles orogéniques proposés pour la Terre primitive, au regard de ceux opérant aujourd'hui. La pétrologie métamorphique, les estimations quantitatives du métamorphisme et celle de la dation U-Pb des monazites métamorphiques mentionnés dans le présent chapitre sont discutés en détail au chapitre 4. Ce chapitre est destiné à être soumis à la revue Precambrian Research début 2018.

Cette étude présente des cartes litho-structurales et métamorphiques nouvellement interprétées sur la ceinture de roches vertes de Sefwi et des domaines volcano-sédimentaires d'âge Paléoprotérozoïque (2300-2070 Ma). Cette ceinture se situe dans la partie sud-ouest du Ghana, marquant la limite orientale du craton ouest-africain. L'utilisation intégrée des observations sur le terrain, la cartographie structurale et l'interprétation des données géophysiques aéroportées régionales fournissent de nouvelles informations sur l'architecture, la séquence de déformation, les relations de superposition tectonique et l'évolution tectonique de la région exposée à l'époque paléoprotérozoïque (2150-2070 Ma). Cette étude met en lumière des domaines tectoniquement juxtaposés, marqués par des grades métamorphiques contrastés, comprenant des roches métamorphisée dans le faciès des schistes verts et amphibolites, des roches volcaniques et volcanoclastiques mafiques à felsiques, des paragneiss de haut grade métamorphique et des ensembles volcano-sédimentaires faiblement métamorphisés. Ces roches para- et ortho-dérivés sont très largement recoupés par plusieurs générations de granitoïdes.

La première déformation se caractérise par une foliation métamorphique omniprésente S1, parallèle à la stratification originelle des roches. Cette foliation précoce est présente sur l'ensemble de la zone d'étude. Elle est plan axial de plis isoclinaux mal préservés Le raccourcissement D1, d'orientation NNO-SSE, est associée à un métamorphisme de haut grade impliquant de la fusion partielle au stade finaux de son évolution. Il est interprété comme lié à un événement d'épaississement de la croûte birimienne, générant des granites à deux micas qui se mettent en place entre 2092 et 2081 Ma, ce qui permet de fixer un âge minimum à cette déformation D1 Les structures prenant place lors de l'événement tectonique ultérieur (D2) sont caractérisées par un pendage moyen, orienté OSO-ENE à NE-SO et un plan de clivage S2 portant une minéralogie métamorphique. La nature transtensive de la déformation est révélée par la nature coaxiale des plis F2 dont les axes sont parallèles à celle de la linéation métamorphique L2, s'orientant ENE-OSO à NE-SO, légèrement oblique par rapport aux grandes zones de cisaillement observées régionalement. Au sein de ce régime transtensif, l'exhumation différentielle tardi-orogénique de la croûte moyenne et inférieure se produit le long des plans extensifs orientés NNE, associant des mouvements en transpression sénestre le long des zones de cisaillement d'échelle régionale orientés NE-SO. La chronologie de l'exhumation est contrainte par des âges métamorphiques U-Pb sur monazites situés autour de ca. 2073 Ma. Le raccourcissement tectonique E-O qui vient se surimposer à l'édifice structurale est interprété comme lié à une modification des conditions aux limites (régionales). Il est associé à la réactivation dextre des zones de cisaillement d'orientation NE-SO et au plissement des structures antérieures, présentes pourtant en position distale par rapport à ces zones de cisaillement majeures. Nous proposons que la limite nord-ouest de la ceinture de Sefwi représente une zone de collision entre d'une part le sud du Ghana et d'autre part, les parties centrales et nord-ouest du Ghana / Côte d'Ivoire. Ces deux domaines présentent des histoires tectoniques contrastées. En outre, nous suggérons que l'accrétion du craton ouest-africain soit le résultat d'une succession d'épisodes d'accrétion au sein de l'orogène éburnéenne, ce qui nous offre de nouvelles perspectives sur l'évolution des processus orogéniques au Paléoprotérozoïque.

Chapitre IV Résumé

Ce chapitre est complémentaire du cadre et de la carte tectono-métamorphiques présentés au chapitre 3, fournissant une analyse pétrographique détaillée, des estimations quantitatives des conditions du métamorphisme régional dominant et des contraintes géochronologiques sur la déformation et le métamorphisme durant l'orogenèse éburnéenne. Ces nouvelles contraintes géochronologiques sont comparées aux données géochronologiques existantes afin de comprendre la chronologie des évènements métamorphiques au sein du craton, mais aussi la

chronologie du métamorphisme de haut grade dans la zone étudiée et son rapport aux autres terrains de haut et bas grades au sein du craton. Les assemblages métamorphiques et les âges sur monazites documentés dans ce chapitre élucident une histoire métamorphique plus complexe pour le sud-ouest du Ghana (en particulier) et l'orogenèse éburnéenne (en générale), contribuant ainsi à une meilleure compréhension de l'amalgamation du craton ouest-africain à cette époque. Cette évolution métamorphique apparait comme un régime d'exception au Paléoprotérozoïque. Ce chapitre a été soumis au Journal of Metamorphic Geology. Une version révisée sera renvoyée avant mai 2018. Les principales révisions proposées ont été prises en compte, certaines révisions étant toujours en cours.

De nouvelles données pétrographiques et géochronologiques sont présentées pour les paragneiss migmatitiques et les roches méta-volcaniques de hauts grades exposées au sud-ouest du Ghana. L'étude porte sur les domaines tectono-métamorphiques présents au cœur de la ceinture de roches vertes de Sefwi et dans les domaines adjacents à cette ceinture. Ces domaines sont bornés ou recoupés par des zones de cisaillement ductiles, fortement réactivées, d'orientation NE à ENE et présentes à l'échelle régionale. Au nord de la ceinture de Sefwi, un terrane de haut grade a été nouvellement identifié. Il comprend des paragneiss migmatitiques hétérogènes de haut grade, injecté par de volumineuses intrusions de granite à deux micas. Il est interprété comme faisant partie de la ceinture de Sefwi en raison de ses similitudes métamorphiques et structurales. Paragneiss et amphibolites à grenat révèlent de façon sporadique un métamorphisme précoce de haute pression et moyenne température (HP-MT) (10,0 à 11,5 kbar, 580-650 ° C), s'inscrivant au sein d'un faciès amphibolite de haute pression et marquant un gradient géothermique apparent assez froid, de l'ordre de 15-17 °C par km. Cet évènement métamorphique précoce évolue vers le facies amphibolite-granulite (7.5 à 9.5 kbars, 650-700 °C) atteignant le début de la fusion partielle (anatexie). Cette évolution se produit le long d'un gradient géothermique apparent moyen de l'ordre de 20-25 °C / km.

La variété des chemins métamorphiques progrades enregistrés par les migmatites révèle un réchauffement durant la décompression, ou inversement un réchauffement durant l'enfouissement. L'exhumation de ces roches partiellement fondues se produit conjointement, comme l'indique le parallélisme entre la forme horaire de tous ces chemins P-T-t, montrant une baisse simultanée en pression et température, et la présence d'un métamorphisme régional dans le facies amphibolite venant se superposer aux assemblages métamorphiques de plus haut grade. Les données d'âge (U-Pb) obtenues i*n-situ* sur monazite (SHRIMP) permettent de situer le début de cette exhumation et de ce refroidissement autour de ca. 2073 Ma. A l'inverse des migmatites, un échantillon d'amphibolite à grenat enregistre son refroidissement durant l'exhumation, sans passer par un stade de plus haute température. Nous interprétons les chemins P-T –t de forme horaire enregistré par deux des trois échantillons, comme la preuve d'un épaississement de la croûte birimienne dans ce secteur de Sefwi, en lien avec la convergence et la collision de ce segment de croûte le long de sa marge nord - ouest. L'exhumation de la croûte inférieure le long des zones de cisaillement d'orientation NE et de plans de détachement à pendage ouest est finalisée aux environs de ca. 2092 - 2081 Ma, ce qui correspond à l'âge de mise en place des leucogranites venant recouper les écailles de croûte inférieure.

Les assemblages métamorphiques de haute pression produits le long de faibles gradients géothermiques apparents distinguent le sud-ouest du Ghana des provinces granitiques-gneissiques précambriennes. De plus, les diverses histoires thermiques et la juxtaposition tectonique de différentes écailles de la croûte issues de différentes profondeurs ne sont pas conformes aux prédictions métamorphiques et au type homogène des déformations prédits par le modèle «d'orogène chaud» proposé pour le cas des orogènes précambriens. De plus, des âges métamorphiques aussi jeunes n'ont pas été documentés auparavant dans la partie sud-est du craton ouest-africain. Ces résultats suggèrent plutôt différents épisodes de collision survenant au cours de l'assemblage du craton, produisant une histoire métamorphique diachrone et complexe. L'enregistrement métamorphique du craton Ouest africain soutient l'idée d'un régime tectonique de *transition* au Paléoprotérozoïque, partagé entre une Tectonique des Plaques moderne qui s'annonce et une tectonique Archaïque qui disparait. Ces gradients métamorphiques froids préfigurent en quelque sorte l'apparition des premières subductions modernes du Néoprotérozoïque et leur cortège métamorphique de ultra-haute pression (UHP).

Chapitre V Résumé

Les chapitres de recherche précédents ont exploré le style tectonique et les mécanismes d'exhumation accompagnant l'orogène éburnéenne, comme il est possible de la documenter dans le SW du Ghana. Ce chapitre se concentre sur l'évolution magmatique de la ceinture de roches vertes de Sefwi afin de mieux comprendre le contexte géodynamique de la croissance de la croûte à cette époque de la Terre. Il présente de nouvelles analyses pétrologiques et

géochimiques des suites magmatiques représentatives de la ceinture de roches vertes de Sefwi et du domaine adjacent de Sunyani-Comoé. L'analyse des données en éléments majeurs et traces permet de remonter aux sources potentielles des magmas et des processus pétrogénétiques responsables de la formation de la croûte birimienne. Les analyses U-Pb et Lu-Hf sur zircons contraignent la chronologie du magmatisme, tout en nous fournissant des informations sur les processus et la chronologie d'extraction de ces magmas à partir du manteau paléoprotérozoïque, mais aussi sur les temps de séjour de ces magmas dans la croûte birimienne et sur la nature de cette proto-croûte en Afrique de l'Ouest. La chronologie du magmatisme et les valeurs $\varepsilon H f_{(t)}$ sont comparés à d'autres régions du craton ouest – africain. Cette approche permet de révéler différents sous-domaines à signature isotopique bien distincte, nous amenant à une meilleure compréhension de l'architecture du craton. Une considération d'ensemble est enfin apportée à cette géodynamique paléoprotérozoïque, responsable de la formation et de la préservation du craton ouest-africain, au regard des modèles géodynamiques proposé pour l'Archéen et ceux connus et opérant au Phanérozoïque.

De nouvelles données géochimiques et isotopiques sont présentées pour les suites magmatiques qui affleurent dans la partie sud-ouest du Ghana (ca. 2300-2070 Ma). Cette étude se concentre sur l'évolution magmatique de la ceinture de roches vertes de Sefwi, redéfinissant les caractéristiques géochimiques et isotopiques des principales suites magmatiques mises en place entre ca. 2155 et 2136 Ma. La croûte paléoprotérozoïque juvénile dérive d'un matériel crustal extrait d'un manteau appauvri entre ca.2650 et 2260 Ma, indiquant des temps de séjour assez courts dans la croûte et impliquant un remaniement très limité de la croûte continentale archéenne présente à cette époque paléoprotérozoïque.

De nouvelles données géochimiques roche totale mettent en lumière la grande variété pétrologique, géochimique et isotopique des suites magmatiques d'âge Paléoprotérozoïque présentes au sud du Ghana. Ces nouvelles données indiquent une mise en place simultanée de ces magmas issus de différentes sources présentent au sein d'arcs volcaniques. Cinq suites magmatiques ont été identifiées: 1) granitoïdes de type tonalites-trondhjémite-granodiorite (TTG), riches en sodium, dominés par la biotite et la hornblende, associés localement à des systèmes effusifs de type dacite; 2) diorites calco-alcalines, enrichies en LILE et laves andésitiques associées; 3) pyroxénites, gabbros et roches intrusives de composition mafique ; 4) monzonites riches en quarz, de composition très potassique; et 5) granites à deux micas et leucogranites à muscovite, enrichis en éléments incompatibles. Les spectres en éléments traces sont tous caractérisés par des valeurs en Nb –Ta, P, Ti très négatives et des anomalies positives en Pb et Ba dans les suites magmatiques plus évolués, ce qui indique une influence de

la croûte au cours de la genèse du magma au sein de l'arc volcanique. Les diorites et les TTGs à biotite présentant de faibles valeurs d'HREE, se mettent en place simultanément vers ca. 2155 Ma, révélant des sources contrastées. Les magmas de type dioritique sont produits plus en profondeur, issus de la fusion partielle d'un coin de manteau métasomatisé et enrichi en éléments incompatibles, tandis que les TTGs à biotite ont probablement été produits par la fusion partielle de panneaux de croûte océanique subductés sous l'arc, au sein d'un manteau assez chaud. La genèse des TTGs à hornblende est attribuée à la fusion partielle à plus faible profondeur d'une croûte mafique. Cette hypothèse se fonde sur l'observation des faibles teneurs en Sr de ces magmas, leur forte teneur en terres rares lourdes (HREE), le faible fractionnement en terres rares légères (REE) et des rapports en éléments assez élevés par rapport aux données issues de l'expérimentale. La mise en place de magma monzonitiques vers ca. 2136 Ma est révélateur d'un processus de mélange au sein de la croûte birimienne, entre des magmas issus de la fusion du manteau et d'autres dérivant d'une refonte partielle de TTGs existants. La mise en place ultérieure de granites à deux micas et des leucogranites, entre ca. 2092-2081 Ma, est interprétée comme le produit d'une anatexie crustale lors d'un évènement de type collision d'arc. La zone de suture envisagée se situerait parallèlement à la marge nord - ouest de la ceinture Sefwi. Les analyses in-situ Lu-Hf sur zircons livrent toujours des valeurs ε Hf_(t) positives, situées entre +4,9 et +8,0 pour les TTG, +2,2 et +5,5 pour les diorites et +2,7 à +5,8 pour les laves dacitiques, mettant en évidence une source nettement radiogénique durant la croissance de la croûte et représentant potentiellement une limite isotopique juvénile au sein de ce craton. Les monzonites mises en place postérieurement révèlent une plus large gamme de valeurs en $Hf_{(t)}$, situées entre +1,2 et +6,5, ce qui suggère une hétérogénéité croissante dans la source de ces magmas. Nous suggérons que la formation et l'accrétion du craton ouest-africain reflètent l'accumulation rapide et l'amalgamation de terranes d'arcs générés en domaine intra-océanique et associant de la subduction. Ce type de géodynamique géochimique est très semblable à celle communément proposée pour les cycles de Wilson et représentait, à ce titre, une de ses toutes premières expression dans l'histoire de la Terre.

Conclusions

La province d'âge Paléoprotérozoïque du Craton ouest-africain fournit un laboratoire naturel pour explorer les liens géodynamiques et orogéniques entre l'Archéen et le Protérozoïque. Ce craton préserve en son coeur un épisode unique de croissance crustale, qui n'a pas d'équivalent

connu au niveau mondial. L'évolution magmatique, tectonique et métamorphique du Craton ouest-africain au Paléoprotérozoïque reflète un régime de Tectonique des Plaques naissant, où s'opère des changements géochimiques et métamorphiques progressifs et majeurs.

Dans le cadre de cette thèse, j'ai établi un nouveau cadre magmatique pour la ceinture de roches vertes de Sefwi au Paléoprotérozoïque, mettant en lumière une production intense de croûte juvénile et la collision d'arcs intra-océaniques au sein d'un régime de subduction. Des similitudes frappantes existent entre les processus pétrogénétiques et l'évolution magmatique des terranes du sud-ouest du Ghana et ceux d'âge Néoarchéen ou de leurs homologues plus modernes, issus des systèmes d'arcs actuels. Les données géochimiques et isotopiques de la région reflètent l'adoption progressive des processus pétrogénétiques et crustaux modernes. Ceci est soutenu par l'histoire tectonique et métamorphique de la région, caractérisée par un épaississement crustal initial en contexte chevauchant puis la génération de conditions métamorphiques de HP-MT, le tout étant interprétée comme la conséquence d'une collision progressive entre terranes d'arc. L'exhumation tectonique ultérieure de la croûte inférieure est la preuve d'un durcissement de la lithosphère sous la province birimienne il y a 2 Ga. Ce comportement mécanique de la lithosphère apparaît incompatible avec le modèle « *d'orogène* chaud » proposé pour le Protérozoïque. L'implication de cette évolution géodynamique et tectonique proposée pour le craton ouest-africain améliore notre compréhension sur l'évolution du régime tectonique global et du refroidissement séculaire de la Terre primitive depuis 2.5 milliards d'années

Sommaire des publications et congrès

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Chapter I

Introduction



1 Introduction

1.1 General Introduction

The application of the plate tectonic paradigm and a uniformitarian approach to interpret the Precambrian Earth remains a contentious issue. Similarities between ancient and modern rock types have prompted many authors to invoke modern plate tectonics by 3000 Ma (Cawood, *et al.*, 2006; Condie & Kröner, 2008; Smithies, *et al.*, 2005; van Hunen & Moyen, 2012), or even earlier (Furnes, *et al.*, 2015), with its operation becoming widespread after 2700 Ma (Condie, 2008). However, a number of the petro-tectonic assemblages and plate tectonic indicators that we use to track the role of plate tectonics, including ophiolite sequences, blueschist facies and ultra-high-pressure metamorphism, did not appear until ca. 1000 Ma (Hamilton, 1998; Stern, 2005; Stern, 2007). Based on this, Stern (2007) suggests that there is a progression from an episodic plate tectonic regime capable of lithosphere recycling and generation of arc-like magmas at ~ 1900 Ma, with sustainable, deep subduction of negatively buoyant oceanic lithosphere beginning in the Neoproterozoic. Alternatively, the global adoption of a plate tectonic regime is suggested to be episodic and diachronous during the Archaean (O'Neill, *et al.*, 2007; van Hunen & Moyen, 2012) or a gradual transitional from a plume-lid regime to a plate tectonic regime from Archaean to the Phanerozoic (Fischer & Gerya, 2016a; Fischer & Gerya, 2016b).

Magmatism and tectonic activity of the Earth are, in essence, the surficial manifestation of the global thermal budget balancing the loss of heat through mantle convection with heat production, primarily generated through radioactive decay (Jaupart, *et al.*, 2007, and references therein). It is widely accepted that the mantle was hotter in the early Earth and has since undergone secular cooling (Korenaga, 2008a; Korenaga, 2008b; Labrosse & Jaupart, 2007; Schubert, *et al.*, 1980), with ambient mantle temperatures at ca. 3000–2500 Ma between 150 and 250 °C higher than present day (Ganne & Feng, 2017; Herzberg, *et al.*, 2010). Elevated mantle temperatures likely affected the composition and thickness of oceanic lithosphere, as well as the rheology, composition and tectonic processes of the continental lithosphere (Burov & Yamato, 2008; Burov, 2011; Davies, 1992; Sizova, *et al.*, 2010; van Hunen & van den Berg, 2008; Yamato, *et al.*, 2008).

In order to understand the evolution of plate tectonics and the continental crust, we turn to the crustal record. This in itself presents geographical and statistical biases, representing a system continually affected by recycling, surface erosion, subduction-zone recycling and overprinting during crustal reworking (Clift & Vannucchi, 2004; Hawkesworth, et al., 2009; Hawkesworth & Kemp, 2006a). Crustal growth models indicate that a significant proportion (~65%) of the continental crust had been extracted from the mantle by the end of the Archaean, with growth rates decreasing after ca. 3000 Ma, coinciding with elevated crustal reworking as indicated by zircon Hf isotopic record (e.g. Belousova, et al., 2010; Dhuime, et al., 2012). Archaean crust, however, represents a relatively low proportion of the continental crust exposed today, reflecting a low preservation rate and intense crustal recycling and reworking in the early Earth, limiting both surface exposure and, consequently, the record of early Earth processes. The igneous and detrital zircon record displays successive peaks, which are interpreted as episodes of discrete juvenile crust production and preservation (Cawood, et al., 2013; Condie, et al., 2011b; Hawkesworth & Kemp, 2006b). The Precambrian geological record shows peak activity at 3500 Ma, 2700 Ma, 2100-1900 Ma and 1100 Ma (Condie, 1998; Condie, et al., 2005; Condie, et al., 2011a; Kemp, et al., 2006; McCulloch & Bennett, 1994) associated with crust production, supercraton/ continent formation and orogenesis. These peaks have multiple interpretations including mantle avalanche events (Condie, 1998) and spikes in super-plume activity (Albarède, 1998; Stein & Hofmann, 1994), or are alternatively correlated with the timing of supercontinent assembly and magmatism (Campbell & Allen, 2008; Taylor & McLennan, 1995). Hawkesworth, et al. (2009), however, note that these peaks may simply represent ideal preservation conditions. Given the ambiguous nature of these crystallisation peaks, exploration of regions that contain significant additions of juvenile crust between these events, such as the ca. 2300 - 2100 Ma crust of the West African Craton (Abouchami, et al., 1990; Boher, et al., 1992), may provide insight into global crustal and tectonic transitions of the early Earth.

There are many changes documented in the magmatic and the metamorphic record that coincide with the Archaean-to-Proterozoic transition, and the hypothesised advent of plate tectonics. Keller and Schoene (2012) describe a significant disruption in the secular evolution of the geochemical characteristics of mafic and felsic lithologies occurring at the Archaean-Proterozoic boundary, which are associated with changes in the degrees of mantle melting, crustal thickness and the rise of atmospheric oxygen and potentially reflect a major change in global geodynamics. MgO-rich komatiites are produced during high degrees of partial melting of a hot mantle (Condie, 1994; Condie & Benn, 2006; De Wit & Ashwal, 1997), and are found in far greater abundance in the Archaean than in either the Proterozoic or the Phanerozoic. Similarly, felsic rocks display strong compositional

variation at the Archaean-Palaeoproterozoic boundary, with Archaean crust dominated by the grey gneisses including the tonalite-trondhjemite-granodiorite (TTG) suite, minor amphibolites and paragneisses. The TTG suite is characterised by high Na₂O and SiO₂ concentrations, high light rare element (LREE) over heavy rare earth element (HREE) ratios and the absence of positive or negative Sr and Eu anomalies (Jahn, et al., 1981; Moyen, 2011; Moyen & Martin, 2012), and are derived from the partial melting of hydrous metabasalts at depth. There are two contrasting schools of thought for their genesis; the first invokes subduction of oceanic crust, often involving young hot lithosphere to transport hydrated basalts to depth (Drummond & Defant, 1990; Halla, et al., 2009; Laurent, et al., 2014; Moyen, 2011), and the second suggests that TTGs are generated during the delamination of an over-thickening mafic crust (Atherton & Petford, 1993; Bédard, 2006; Smithies, 2000). Magmatism in the Neoarchaean reveals greater petrological and geochemical diversity with the emplacement of TTGs followed by high-K granitoids, two mica-granites and late sanukitoids (Laurent, et al., 2014; Martin, et al., 2009; Zeh, et al., 2009). The sanukitoids series have potassic calc-alkaline to alkalicalcic, metaluminous compositions and are enriched in both incompatible elements, and Ni and Cr, suggesting both crustal and mantle-derived contributions, or contamination during melting from basic to ultrabasic rocks (Laurent, et al., 2014; Martin, et al., 2009). As such, they are interpreted as the product of the interaction of melts derived from enriched crustal material and mantle peridotite, and are therefore a proposed proxy for the development of subduction-related processes at the Archaean-Proterozoic boundary (Heilimo, et al., 2010; Laurent, et al., 2011; Moyen, et al., 2003).

Phanerozoic versus Precambrian orogenesis

In order to understand the development or onset of the plate tectonic regime, many geologists have long relied upon uniformitarianism, interpreting features of ancient terranes as analogues to modern geological formation (Cawood, *et al.*, 2009; Condie & Benn, 2006; Condie & Kröner, 2008; Ernst, 2009; Stern, 2005; Windley, 1981). Alternatively, based on archetypal features of Archaean provinces, authors have invoked unique geodynamic settings for crustal evolution during the early Earth (Bédard, *et al.*, 2003; Bédard, *et al.*, 2013; Chardon, *et al.*, 1998; Hamilton, 1998; Smithies, *et al.*, 2003; Van Kranendonk, *et al.*, 2007). Here we explore the characteristics and tectonic differences of Phanerozoic, Proterozoic and Archaean orogens, in order to discuss their geodynamic significance and the contrasting schools of thought regarding Precambrian tectonics.

The modern plate tectonic paradigm is largely based on our understanding and documentation of Phanerozoic orogens that form along convergent plate margins in association with the subduction of oceanic lithosphere. Convergence and lateral shortening are responsible for the accretion of mobile orogenic belts or terranes with contrasting tectono-metamorphic histories. The horizontal forces generated by subduction lead to the centripetal growth of continental crust, evidenced by the amalgamation of younger terranes on continental margins and lateral age gradients (Moresi, et al., 2014; Percival, et al., 2006). In general, Phanerozoic accretionary orogens comprise accretionary wedges, exotic crustal fragments, volcanic arcs and associated back arcs, ophiolites, clastic sedimentary basins and distinct metamorphic terranes (Cawood, et al., 2009; Condie, 2005; Hyndman, et al., 2005). The forearc is characterised by low heat flow decreasing in a trenchward direction with the downgoing cold lithosphere recording a low geothermal gradient of 5-10 °C/km, characteristic of a "cold" orogen. In contrast, the magmatic arc and the backarc domains are characterised by high heat flow and high geothermal gradients between 20 and 40 °C/km (Hyndman, et al., 2005). The concurrent burial and metamorphism of the accretionary prism and rocks of the arc and back-arc domains under blueschist-to-eclogite facies metamorphic conditions and amphibolite-to-granulitefacies metamorphic conditions, respectively, and their subsequent juxtaposition, creates a thermal asymmetry within the accretionary orogenic systems. The duality of thermal regimes represents a hallmark indicator of modern accretionary orogens generated in a plate tectonic regime (Brown, 2006).

Continued convergence results in the closure of oceanic basins, the onset of collision and the tectonic assembly of domains within the accretionary orogen or exisiting continental lithosphere, exemplified by the Alpine-Himalayan Orogen. In this setting, lateral shortening is accommodated by the development of major thrust faults, nappe stacking and heterogeneous crustal thickening within the accretionary wedge and foreland belts and basins (Dahlen, 1990). The resulting elevated topographic profiles are continually subjected to erosional and tectonic processes that ultimately remove the overburden, exposing the high-pressure rocks from deep within the orogen that yield key information regarding the tectonic processes of Phanerozoic collisional orogens. First documented in the Phanerozoic Western Alps continental collisional belt by Chopin (1984) and in the ca. 400 Ma Caledonides of Norway (Smith, 1984), coesite inclusions in garnet and clinopyroxene porphyroblasts within high-grade blueschist domain and eclogite pods revealed the burial of continental upper crustal rocks within the subduction zone to depths of approximately 100 km (~ 30 kbar) and their subsequent

return to the surface (Chopin, 1984; Chopin, 2003; Smith, 1984). The interpreted mechanism for the burial involves the entrainment of tracts of the leading edge of a continental plate, attributed to body forces during incipient collision (Chopin, 2003, and references therein). Greater debate, however, surrounds the mechanism for the exhumation of ultra-high pressure (UHP) rocks. Petrological and isotopic data reveal simultaneous high cooling rates and high decompressions rates, representative of rapid exhumation rates of 1-2 cm/yr (Burov, et al., 2001; Doin & Henry, 2001; Hacker, et al., 2000), with observed concurrent cooling during exhumation (Carswell & Zhang, 1999; Chopin, 1984) precluding entrainment of UHP metamorphic rocks during the diapiric ascent of plutons as a viable exhumation mechanism. Burov, et al. (2001, and references therein) summarised multiple levels of exhumation in a continental collision zone. They propose that exhumation of low-pressure-lowtemperature rocks of the accretionary prism can be attributed to classical corner flow, whilst deeper high-pressure-high-temperature segments of the crust are exhumed through positive buoyancy and ascending shear flow. Alternatively, exhumation of the deepest material is attributed to convective instabilities generated through heating of crustal material by the overlying asthenospheric mantle (Burov, et al., 2001), or to the break off subducted lithosphere and its subsequent exhumation along the subduction channel due to its relative positive buoyancy (Gerya, et al., 2002). These mechanisms are then complemented by the removal of overburden and thinning of the crust during extensional tectonics. Extension is attributed to extensional collapse (Andersen, 1998), lateral, ductile flow of a weak, partially molten lower crust (Vanderhaeghe & Teyssier, 2001a; Vanderhaeghe & Teyssier, 2001b) or, alternatively, to slab rollback (Brun & Faccenna, 2008), ultimately exposing the complex melange of metamorphic terranes observed in orogenic belts today.

Numerical experiments and thermomechanical modelling (Burov, *et al.*, 2014; Burov & Yamato, 2008; Faccenda, *et al.*, 2008; Yamato, *et al.*, 2007) examining the role and interaction of these processes are complemented and constrained by petrological evidence and the determination of *P-T-t-d* paths of metamorphic terranes (e.g. England & Thompson, 1984), which allow us to calculate transient geotherms within the crust during orogenesis. This knowledge often guides interpretation of early Earth tectonics or alternatively highlights the differences in architecture, processes and geotherms that characterise Archaean and Palaeoproterozoic provinces. The following section explores the unique characteristics of Precambrian orogens and the range of proposed geodynamic frameworks, including ambiguity of evidence between different end-member models.
Archaean provinces, and and some Palaeoproterozoic terranes, frequently display a unique crustal architecture characterised by narrow greenstone belts, vast granite-gneiss terranes dominated by TTG magmatism and craton-scale shear zones. The interpreted geodynamic significance of such terranes follows two schools of thought, the first of which regards granite-greenstone provinces as ancient analogues of the modern, horizontal, subduction-driven plate tectonic regime (Cawood, et al., 2006; Condie & Benn, 2013; Condie & Kröner, 2008). In this model, crustal-scale shear zone are interpreted as major thrust related sutures. The second school of thought invokes a unique geodynamic model for the early Earth (Bédard, 2003; Bédard, et al., 2013; Smithies, et al., 2003). Archaean metamorphism predominantly preserves non-linear exposures of greenschist to mediumpressure granulite facies rocks, with no record of blueschist facies metamorphism, nor deep continental subduction (Brown, 2007, and references therein). Strain patterns in some Archaean provinces define dome and keel geometries with strain trajectory triple junctions associated with homogeneous distributed deformation. These strain patterns are interpreted as the product of both body forces exerted by diapiric structures and regional horizontal shortening reflecting boundary forces (Bouhallier, et al., 1995; Choukroune, et al., 1995). Gravitational instabilities between the thick, dense overlying (flood) basalts or greenstone cover and the underlying warm, felsic and less dense granitoid-dominant lower crust result in a density inversion (Abbott & Hoffman, 1984; Chardon, et al., 1996; Chardon, et al., 1998; Hoffman & Ranalli, 1988; Taylor & McLennan, 1985; Van Kranendonk, et al., 2004; West & Mareschal, 1979). Synformal structures commonly observed in anastomosing greenstone belts are attributed to the downward advection or sagduction of cold dense material, and elliptical foliation patterns form during the diapiric emplacement of high-viscosity granitoids (Chardon, et al., 2002; Choukroune, et al., 1995). The dome-and-keel geometries contain vertical stretching lineations at the contact between the two domains and are thus interpreted as the result of "vertical" tectonics (Bouhallier, et al., 1995; Choukroune, et al., 1995; Lompo, 2010; Marshak, 1999; Van Kranendonk, et *al.*, 2004; Vidal, *et al.*, 2009).

However, the dichotomic debate between vertical and horizontal tectonics has recently been questioned by a number of authors (François, *et al.*, 2014; Rey, *et al.*, 2014; Thébaud & Rey, 2013). François, *et al.* (2014) integrated structural, metamorphic and geochronological studies in the Archean East Pilbara granite-greenstone terranes, which revealed the rapid burial (9–11 kbar, 450–550 °C) of metabasalts and metasediments and their subsequent exhumation during partial melting of the lower crustal and the formation of granitoid domes. Complementary thermomechanical modelling

revealed a diverse range of apparent geothermal gradients (10-45°C/km) (François, et al., 2014). In addition, based on numerical experiments, Thébaud and Rey (2013) argue that sagduction can generate significant lateral shortening and intraplate deformation. The rapid burial of supracrustal sequences and associated cold apparent geothermal gradients are typically attributed to subduction along a plate boundary, however, both François, et al. (2014) and Thébaud and Rey (2013) highlight the potential of sagduction as a viable alternative mechanism for burial and exhumation. A similar mechamism was proposed by Ganne, et al., (2014) for high-pressure (>10 kbar) volcano-sedimentary Palaeoproterozoic rocks of the West African Craton, which were exhumated from lower crustal depths through entrainment in ascending granitoids. It was recented noted, however, by Rey, et al. (2014) that sagduction and subduction are not neccessarily mutally exclusive when dealing with elevated amntle temperatures. Numerical modelling suggests that early continents derived from differentiation of thick oceanic plateaus underwent intraplate gravitational collapse, which generated sufficient lateral forces and buoyancy contrasts to initiate episodic subduction of adjacent oceanic lids. Using this method, Rey, et al., (2014) were able to generate a self-consistent geodynamic framework explaining multimodal volcanism and TTG melts and formation of a deep mantle root, characteristic of Archaean cratons, as well as coexisting stagnant-lid and subduction-like processes. Given these arguments, distinguishing between vertical and horizontal tectonics and different geodynamic settings on the basis of deformation and apparent geothermal gradients is remains problematic and nuanced.

The third end-member orogenic model invoked in the Precambrian tectonic debate is the "hot-orogen" model proposed for young, hot Precambrian orogens, sourcing evidence from field observations, strain patterns and analogue and numberical models. In this model, bulk horizontal contraction is accommodated by homogeneous deformation and metamorphism and lateral crustal flow due to rheologically weakened lithosphere (Chardon, *et al.*, 2009; Chardon, *et al.*, 2011; Cruden, *et al.*, 2006; Rey & Houseman, 2006). The rheologically weak lithosphere prohibits the formation of both significant topographic relief and crustal-scale detachments capable of exhuming the lower crust. The hot-orogen mode is invoked for Precambrian accretionary and collisional orogens (e.g. Cagnard, *et al.*, 2007; Vidal, *et al.*, 2009) as well as wide, mature Phanerozoic orogens such as the Variscan Belt (Schulmann, *et al.*, 2008).

Whilst the debate surrounding early Earth tectonics and geodynamics continues to focus on finite timing and end-member models, other datasets suggest a more gradual progression towards the

modern plate tectonic regime. In the metamorphic record, Archaean metamorphic rocks commonly exhibit average metamorphic conditions from sub-greenschist facies to granulite facies with no sharp lateral metamorphic breaks (e.g. Goscombe, *et al.*, 2009; Komiya, *et al.*, 2002; Percival, *et al.*, 1992). The Neoarchaean was marked by the coeval appearance of ultra-high temperature (UHT) and eclogite-high-pressure granulite (E-HPG) facies metamorphism, interpreted by Brown (2006; 2007) as a key marker for the onset of a *transitional* Proterozoic plate tectonic regime involving increasingly widespread subduction-like process transitioning to self-sustaining subduction systems around the globe. It was not until the Neoproterozoic that the classic metamorphic indicators of the modern plate tectonic regime appearance including blueschist facies metamorphic conditions and ultra-high pressure assemblages (Stern, 2005; Stern, 2007), suggesting the progressive evolution and adoption of present day subduction dynamics.

Many datasets derived from the geological record suggest the Earth was going through significant geodynamic changes at the Archaean-Palaeoproterozoic boundary, which possibly extended into the Palaeoproterozoic (e.g. Brown, 2007; Condie, 1997; Keller & Schoene, 2012). Given the breadth of time for which estimates of the commencement of plate tectonics span (~700 Ma to >4000 Ma: Harrison, *et al.*, 2005; Sleep, 2005; Stern, 2007; van Hunen & Moyen, 2012), a focus on the geological record during this transitional period is warranted. The Palaeoproterozoic provinces of the West African Craton are characterised by juvenile crust composed of volcano-plutonic belts and granite-gneiss domains similar to Archaean granite-greenstone provinces (Abouchami, *et al.*, 1990; Lompo, 2010; Milési, *et al.*, 1992; Taylor, *et al.*, 1992; Vidal, *et al.*, 2009).The craton represents a significant crustal growth event between the two global peaks in juvenile crust production and the first orogenic event during a period of global orogenesis between 2100 and 1800 Ma, heralding the amalgamation of the first Palaeoproterozoic supercontinent, Nuna (Columbia) (Rogers & Santosh, 2002; Zhao, *et al.*, 2002; Zhao, *et al.*, 2004). As such, the West African Craton provides a unique opportunity to explore the enigmatic tectonic regime of the Proterozoic.

1.2 West African Craton Geology

1.2.1 Introduction

The Palaeoproterozoic (2300–2070 Ma) West African Craton represents one of Earth's oldest Palaeoproterozoic accretionary orogens (Windley, 1992). This orogeny is associated with high



Fig. 2.1. Simplified geological map of the West African Craton, showing key geological provinces (modified after Ennih & Liégeois, 2008).

production rates of juvenile crust and extensive, protracted magmatism (Abouchami, *et al.*, 1990; Condie, 2007), forming between two global crustal growth maxima at 2700 and 1900 Ma (Cawood, *et al., 2*013; Condie, 1998; Condie, *et al., 2*005). The formation and deformation of the craton precedes the amalgamation of the first Palaeoproterozoic supercontinent, Nuna (Columbia) and a major period of global orogenesis between 2100 and 1800 Ma (Rogers & Santosh, 2004; Zhao, *et al.*, 2002; Zhao, *et al.*, 2004). The following section provides the geological setting of the West African Craton, highlighting controversies that persist in the literature, reminiscent of those that arise in the Neoarchaean record.

1.2.2 Geological Setting

The West African Craton consists of two major geological provinces: the Reguibat Rise of northwest Africa and the Leo-Man Rise of western Africa (Fig. 2.1) (Bessoles, 1977b; Black, 1980; Rocci, *et al.*, 1991). The Reguibat Rise extends across southern Morocco to southern Algeria, whilst the Leo-Man Rise is exposed throughout Sierra Leone, Liberia, Guinea, Ivory Coast, southern Mali, Burkina Faso, Ghana and Niger, and includes the Kédougou-Kéniéba and Keyes inliers of Senegal and western Mali. The craton is centrally overlain by the intracontinental Mesoproterozoic to Devonian Taoudeni Basin and the Neoproterozoic Volta Basin in the southeast. The entire craton is bounded to the east and west by Pan African-aged mobile belts (Ennih & Liégeois, 2008; Schluter & Trauth, 2008; Villeneuve & Cornée, 1994). Prior to the opening of the Atlantic Ocean in the Jurassic, correlations are made between the West African Craton, the São-Luís Craton and the north-eastern extent of French Guiana (Klein, *et al.*, 2005; Klein & Moura, 2008; Ledru, *et al.*, 1994; Nomade, *et al.*, 2003; Vanderhaeghe, *et al.*, 1998).

Both the Reguibat Rise and the Leo-Man Rise each comprise an Archaean domain in the west and a Palaeoproterozoic domain to the east, separated by major shear zones (Fig. 2.1) (Black, 1980). The Archaean nuclei of the Leo-Man Rise comprises the Kenema-Man Domain, consisting of tonalite-trondhjemite-granodiorite gneisses emplaced in three discrete pulses at 3600–3500 Ma, 3260–3050 Ma and 2960–2850 Ma, intercalated with linear belts of supracrustal rocks including basalts, BIFs, komatiites and minor sedimentary sequences (Rollinson, 2016, and references therein). Crust formation in the Leo-Man Rise is concomitant with the magmatism of the Reguibat Rise (Potrel, *et al.*, 1998; Potrel, *et al.*, 1996). The Kénéma-Man Domain was affected by two Archaean orogenic events, including the Leonian (3500–2900 Ma) and the Liberian (2900–2600 Ma) events (Bessoles, 1977b; Rollinson, 2016; Tagini, 1971a), associated with amphibolite- to granulite facies regional metamorphism (Kouamelan, *et al.*, 1998; Kouamelan, *et al.*, 1997; Thiéblemont, *et al.*, 2001). The final orogenic event is attributed to the accretion of the Palaeoproterozoic domains during the Eburnean Orogeny (ca. 2100 Ma) (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992; Liégeois, *et al.*, 1991).

Palaeoproterozoic rocks of the West African Craton form the eastern domains of the

Reguibat and Leo-Man Rises, known as the Yetti-Eglab and Baoulé-Mossi Domains, respectively. The Palaeoproterozoic Baoulé-Mossi domain of the West African Craton (WAC) is characterised by elongate meta-volcanic greenstone belts, meta-sedimentary basins, and large granite-gneiss provinces consisting of variably deformed grey gneisses and tonalite-trondhjemite-granodiorite (TTG) suites, separated by craton-scale shear zones (e.g. Abouchami, et al., 1990; Bessoles, 1977b; Feybesse & Milési, 1994). Volcanic and volcano-sedimentary sequences for the Birimian Supergroup were first described in the Birim Valley in Ghana by Kitson (1918). Greenstone belts predominantly comprise metamorphosed Birimian volcanic sequences, which range in composition from tholeiitic and calcalkaline basalts through to rhyolites and associated minor volcaniclastic, greywacke and chemical sedimentary sequences (Baratoux, et al., 2011; Boher, et al., 1992; Doumbia, et al., 1998; Pouclet, et al., 2006). Birimian volcano-sedimentary rocks are made of low-grade argillite, shale, greywacke and rare intercalated intermediate volcanoclastic or volcanic layers (Adadey, et al., 2009; Agyei Duodu, et al., 2009; Junner, 1940). The youngest Palaeoproterozoic sequence comprises fluvio-deltaic detrital sedimentary formations of the Tarkwa Group, which are often spatially restricted to fault bounded basins overlying central portions of greenstone belts (Ledru, et al., 1994). Tarkwaian rocks consist of conglomerates, quartzites and immature sediments, defined by the sequences exposed in the Ashanti Belt in southern Ghana (Junner, 1935; Perrouty, et al., 2012). Timing of deposition was originally constrained by detrital zircons from the basal units of the formation yielding single U-Pb and Pb-Pb ages between ca. 2150 and 2143 Ma (Davis, et al., 1994; Hirdes & Nunoo, 1994). The maximum age of deposition has subsequently been revised to 2107 - 2097 Ma (Perrouty, et al., 2012; Pigois, et al., 2003). Tarkwa Group sedimentary rocks are interpreted as syn-orogenic to late stage basins (Davis, et al., 1994; Perrouty, et al., 2012), or alternatively as a molasses sequence deposited at the onset of deformation in a foreland basin (Eisenlohr & Hirdes, 1992).

Craton-scale isotopic studies using whole rock Nd and Lu-Hf analysis of magmatic and detrital zircons predominantly indicated a radiogenic source for the Palaeoproterozoic crust (Abati, *et al.*, 2012; Abouchami, *et al.*, 1990; Block, *et al.*, 2016a; Boher, *et al.*, 1992; Klein, *et al.*, 2005; Tapsoba, *et al.*, 2013). Hf-and Nd-isotope signatures yield model ages on average between 2500 and 2100 Ma, indicating short crustal residence after mantle extraction. A number of studies, however, reveal negative $\varepsilon Nd_{(t)}$ and $\varepsilon Hf_{(t)}$ values in far south-eastern Ghana and southern Mali, indicating an older crustal influence or reworking of Archaean crust, with model age ranging from 4000 to 2800 Ma in southern Mali (Parra-Avila, *et al.*, 2016; Petersson, *et al.*, 2016; Taylor, *et al.*, 1992). The oldest

detrital zircons in the Palaeoproterozoic domain yield U-Pb ages between ca. 3600 and 2700 Ma sourced from southern Mali (Parra-Avila, *et al.*, 2016). These ages correlate well with the Kénéma-Man Domain, yet are not found elsewhere in the craton. Inherited zircon cores from throughout the domain predominantly yield ages between ca. 2300 and 2200 Ma (de Kock, *et al.*, 2011; Gasquet, *et al.*, 2003; Siegfried, *et al.*, 2009; Thomas, *et al.*, 2009), whilst the oldest recorded crystallisation ages are yielded by granodiorite gneisses in Burkina Faso at 2265 ± 17 Ma (Parra-Avila, *et al.*, 2017) and 2253 ± 9 Ma (Tshibubudze, *et al.*, 2013).

Bimodal volcanism of the Birimian greenstone belts is dated across the craton at between 2300 and 2180 Ma (Block, *et al.*, 2016a; Delor, *et al.*, 1995; Feybesse, *et al.*, 2006; Hirdes & Davis, 1998; Hirdes, *et al.*, 1996; Hirdes, *et al.*, 2007). Similar age ranges are yielded by the earliest phase of TTG and calc-alkaline plutonism in Ghana (de Kock, *et al.*, 2011; Doumbia, *et al.*, 1998; Feybesse, *et al.*, 2006; Siegfried, *et al.*, 2009), Burkina Faso (Parra-Avila, *et al.*, 2016; Tshibubudze, *et al.*, 2015), Ivory Coast (Gasquet, *et al.*, 2003; Pouclet, *et al.*, 2006) and Senegal (Dioh, *et al.*, 2006; Gueye, *et al.*, 2008; Gueye, *et al.*, 2007), with granitoid emplacement occurring near continuously until 2100 Ma. Post-2100 Ma volcanism and calc-alkaline plutonism is documented in Kedougou-Kéniéba Inlier of Senegal (Hirdes & Davis, 2002), prompting authors to propose a diachronous accretion of the Palaeoproterozoic crust; however, magmatism in the remainder of the craton at this time is limited to the emplacement of late potassic granites and leucogranites along the Palaeoproterozoic-Archaean boundary and minor occurrences in southern Ghana (Egal, *et al.*, 2002; Gueye, *et al.*, 2007; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007).

Deformation and metamorphism of the Palaeoproterozoic domains is attributed to the Eburnean Orogeny (2150 – 2070 Ma) (Bonhomme, 1962; Feybesse, *et al.*, 2006). Eburnean deformation is characterised as an early crustal thickening event, often associated with extensive magmatism and greenschist to granulite facies metamorphism, followed by the development of craton-scale shear zones during transcurrent tectonism (Baratoux, *et al.*, 2011; Eisenlohr & Hirdes, 1992; Feybesse, *et al.*, 2006; Hein, 2010; Jessell, *et al.*, 2012; Perrouty, *et al.*, 2012; Tagini, 1971a; Vidal, *et al.*, 2009). The early phase of deformation was responsible for juxtaposition of domains of different composition and grade and the generation of high-grade terranes, which authors attribute to a range of mechanisms, discussed in detail in Section 2.3. Major strike-faulting in the later stages of the Eburnean (ca. 2100 Ma) is commonly associated with a pervasive greenschist facies overprint and wide-spread gold mineralisation hosted in shear zones and palaeoplacer deposits (Feybesse, *et al.*, 2006; Galipp, *et al.*, 2003; Ledru, *et al.*, 1991;

Milesi, *et al.*, 1989; Oberthür, *et al.*, 1998; Pigois, *et al.*, 2003; Vidal, *et al.*, 2009). Terminal collision or docking of the Palaeoproterozoic and Archaean domains is constrained by formation and exhumation of high-pressure granulite facies rocks between 2050 and 2030 Ma (Kouamelan, 1996; Kouamelan, *et al.*, 1997; Pitra, *et al.*, 2010). The craton records very few magmatic ages after ca. 1980 Ma, indicating a prolonged period of quiescence until ca. 1000 Ma (Ennih & Liégeois, 2008, and reference therein). This dramatically contrasts the global crustal and orogenic record, which is dominated by the elevated crustal growth and widespread formation of linear, mobile orogenic belts during the formation and assembly of the supercontinent, Nuna (Condie, 1998; Zhao, *et al.*, 2002).

1.2.3 Controversies

1.2.3.1 Stratigraphy and nomenclature

The Palaeoproterozoic West African Craton is characterised by narrow volcanic greenstone belts, wide meta-sedimentary "basins" and vast granite-gneiss terranes, bounded and separated by crustal-scale transcurrent shear zones that formed during the Eburnean Orogeny. Sporadic exposures of high metamorphic grade gneissic to migmatitic rocks occur throughout the Baoulé-Mossi domain, juxtaposed with low-grade rocks. Questions that continue to arise include: What is the tectono-stratigraphic architecture of the craton? Can we correlate stratigraphic units and deformation events at a craton-scale? Was the Eburnean Orogeny monocyclic or polycyclic? What do the strain patterns and metamorphic record of the craton tell us about the tectonic style during the Eburnean?

Previous studies have employed geochemical, geochronological and structural investigations in order to determine the stratigraphic framework, the tectonic environment and the phases and styles of tectonism recorded within the craton. Persisting controversies and questions regarding the Palae-oproterozoic evolution of the craton are mainly due to the limited geochemical, metamorphic and geochronological data, and the integrations of such datasets at a regional or craton scale, as well as complex geo-political relationships. This has produced a variety of evolutionary models for the West African Craton (Abouchami, *et al.*, 1990; Baratoux, *et al.*, 2011; Block, *et al.*, 2016b; Boher, *et al.*, 1992; de Kock, *et al.*, 2012; Egal, *et al.*, 2002; Feybesse, *et al.*, 2006; Feybesse & Milési, 1994; Hein, 2010; Ledru, *et al.*, 1994; Liégeois, *et al.*, 1991; Milesi, *et al.*, 1989; Tshibubudze, *et al.*, 2009; Vidal & Alric, 1994; Vidal, *et al.*, 2009).

Some of the earliest controversies generated in literature regarding the West African Craton, arose

from conflicting interpretation of the stratigraphy of the craton between French and English geologists in the 1940 – 1970s. Early literature, predominantly proposed by Anglophone geologists, divided the Birimian Supergroup into a lower series of greywackes, phyllites, shales and minor lavas, and an upper series of mafic to intermediate lavas and subordinate volcaniclastic rocks, greywacke and phyllite (Junner, 1935; Junner, 1940; Kesse, 1985; Ledru, et al., 1989; Lemoine, et al., 1985; Milési, et al., 1986). Conversely, francophone geologists proposed a basal Birimian volcanic and volcano-sedimentary unit, overlain by shallow basins filled with flysche-type and molasse sedimentary sequences (Aubouin, 1961; Bessoles, 1977b; Tagini, 1971b). In addition, structural, geochemical and isotopic studies by Eisenlohr and Hirdes (1992) and Leube, et al. (1990) suggest that the Birimian metavolcanic rocks and metasedimentary formations are contemporaneous distal, lateral facies equivalents. Most recently, Adadey, et al. (2009) and Perrouty, et al. (2012) proposed a stratigraphy for the Birimian Supergroup in southern Ghana using a compilation of the growing geochronological data for the region. The stratigraphy comprises the basal Sefwi Group metavolcanics and mica schists (>2174 Ma), constrained by the syn-tectonic emplacement of the Sekondi granitoid (Oberthür, et al., 1998). The overlying Kumasi Group phyllites and volcanoclastic rocks were deposited between ca. 2154-2125 Ma, based on detrital zircon ages, syn-depositional volcanism and late-depositional granitoid intrusion (Adadey, et al., 2009; Oberthür, et al., 1998; Perrouty, et al., 2012).

Whilst the advent of precise geochronology has largely settled the stratigraphic debate, there are contrasting interpretations of deformation relationships. This includes suggestions of an early deformation event recorded by the lower meta-volcanic units that is not recognised in upper meta-sedimentary units, which suggests a polycyclic evolution for the Eburnean Orogeny (de Kock, *et al.*, 2012; Feybesse, *et al.*, 1990; Ledru, *et al.*, 1991; Perrouty, *et al.*, 2012). This model suggests that there was an early orogenic event prior to the emplacement of the Birimian meta-sedimentary rocks. Authors suggest that high-grade terranes and some greenstone belts record elevated metamorphic conditions and polyphase deformation, interpreted as an older basement upon which meta-sedimentary units were deposited and subsequently deformed during the Eburnean Orogeny *sensu stricto (s.s.)* (e.g. Arnould, 1961; Lemoine, *et al.*, 1990; Perrouty, *et al.*, 2012). The age and terminology of the early deformation event differs throughout the Baoulé-Mossi Domain, including the "Burkinian" orogenic cycle in the Ivory Coast (2400 – 2150 Ma) (Lemoine, *et al.*, 1990), the "Tangaean" event of Burkina Faso (2170 – 2130 Ma) (Hein, 2010; Tshibubudze, *et al.*, 2009), or, more recently, "Eburnean I" (2266 – 2150 Ma) (Allibone, *et al.*, 2002) or "Eoeburnean" event in Ghana (2195 – 2150 Ma) (Baratoux, *et al.*, 2011; de Kock, *et al.*, 2011; de Kock, *et al.*, 2012; Perrouty, *et al.*, 2012). Early deformation often corresponds with an initial phase of magmatic accretion and crustal thickening (e.g. Feybesse, *et al.*, 2006; Milési, *et al.*, 1992; Vidal, *et al.*, 2009). Each of these events are interpreted to precede the tectonic assembly of the Paleoproterozoic and Archaean domains and transcurrent tectonism and greenschist facies meta-morphism attributed to the Eburnean Orogeny *s.s.* (2130 – 1980 Ma) (Feybesse, *et al.*, 2006; Jessell, *et al.*, 2012; Ledru, *et al.*, 1991; Lompo, 2010; Oberthür, *et al.*, 1998; Pitra, *et al.*, 2010).

Monocyclic interpretations for the Eburnean Orogeny deem volcano-plutonic belts and metasedimentary basins to be contemporaneous, lateral facies equivalents or representative of crustal segments of varying depths. This interpretation requires greenschist facies metasedimentary provinces to be coeval, supracrustal equivalents to high grade metamorphic terranes, which were juxtaposed during later deformation (Block, *et al.*, 2015; Block, *et al.*, 2016b; Eisenlohr & Hirdes, 1992; Hirdes, *et al.*, 1996; Hirdes, *et al.*, 2007; Leube, *et al.*, 1990; Opare-Addo, *et al.*, 1993). These models suggest that deformation and metamorphism across the craton reflect a long-lived, progressive orogenic event (Block, *et al.*, 2016b; Eisenlohr & Hirdes, 1992; Hirdes, *et al.*, 2003). Given the variety of interpretations and discrepancies between the terranes, additional geochronologically-constrained structural and metamorphic studies are required in order to elucidate the spatial and temporal variations of deformation events across the craton.

1.2.3.2 Metamorphic record and tectonic style of the West African Craton

Metamorphism of the meta-volcanic and meta-sedimentary rocks within greenstone belts reveal predominantly low-grade metamorphism from greenschist to amphibolite facies (Galipp, *et al.*, 2003; John, *et al.*, 1999; Kříbek, *et al.*, 2008). In contrast, metamorphism and migmatization of orthogneisses and paragneisses in high-grade terranes is associated with upper amphibolite to granulite facies conditions (Bessoles, 1977a; Block, *et al.*, 2015; de Kock, *et al.*, 2011; John, *et al.*, 1999; Opare-Addo, *et al.*, 1993), with high-pressure granulites predominantly restricted to the tectonic contact between the Archaean Kénéma-Man and Palaeoproterozoic Baoulé-Mossi domains (Pitra, *et al.*, 2010). Whilst geochronological data is relatively limited, the timing of Eburnean metamorphism is constrained by U-Pb ages of metamorphic monazite, zircon and titanite in between 2110 and 2080 Ma (e.g. de Kock, *et al.*, 2011; Oberthür, *et al.*, 1998), with older paroxysmal metamorphism at ca. 2130 Ma proposed by Block, *et al.* (2015; 2016b) based on metamorphic monazite and zircon ages. Sm-Nd garnet-whole rock isochron ages from central Ivory Coast are older still at ca. 2150 Ma (Boher, *et al.*, 1992). Analyses of strain patterns and metamorphic relationships have drawn a range of interpretations on the tectonic styles and geodynamic mechanisms responsible for the tectonic evolution of the West African Craton. Some authors interpret dome and basin geometries within the craton, which they suggest was associated with Archaean-like diapirism-related vertical tectonics, with late horizontal shortening purely accommodated by transcurrent shear systems (e.g. Lompo, 2010; Pouclet, et al., 2006; Pouclet, et al., 1996; Vidal, et al., 1996). In such studies, amphibolite facies metamorphism is attributed to contact metamorphism during granitoid emplacement (Debat, et al., 2003; Gasquet, et al., 2003; Pons, et al., 1995; Soumaila & Garba, 2006). In the case of the Sefwi Greenstone Belt, large tracts of regional amphibolite facies gneisses are peculiarly attributed to an underlying, unexposed K-rich pluton (Hirdes, et al., 1993). Alternative proposals for orogenic styles in West Africa include "modern-type" thrust-related tectonics and nappe stacking (Feybesse, et al., 2006; Feybesse & Milési, 1994; Milési, et al., 1992). Indeed, the cold (~15°C/km) apparent geothermal gradients deduced for rocks metamorphosed at the greenschist-blueschist facies transition are interpreted as evidence of modern-style subduction during the formation and assembly of the craton (Ganne, et al., 2012). Preservation of cold geothermal gradients is inconsistent with many studies which interpret deformation and metamorphism of the West African Craton as a weak, hot orogen (Caby, et al., 2000; Lompo, 2009; Pons, et al., 1995; Vidal, et al., 2009). The weak, hot orogen model consists of extensive magmatism, often diapirically emplaced, with subsequent lateral shortening associated with homogeneous thickening and late-orogenic development of large-scale shear zones. The high-pressure granulite facies metamorphism at the Archaean-Palaeoproterozoic boundary along the Sassandra Shear Zone in western Ivory Coast may be elucidated by either crustal thickening by unit stacking in a collisional setting or homogeneous thickening of a warm, weak crust (Pitra, et al., 2010). In NW Ghana, late-orogenic exhumation of amphibolite-granulite facies rocks occurs during orogenic collapse following crustal thickening, with the formation of anatectic domes resembling modern collisional orogens (Block, et al., 2016b). Given the diversity of tectonic models, deformation styles and distribution of metamorphic terranes proposed for the West African Craton, it is possible that the craton preserved spatial and temporal variations in the crust. Elucidation of such variations requires continued integration of structural, metamorphic and geochronological studies.

1.2.3.3 Geodynamic setting and crustal formation

Based on $\varepsilon Nd_{(t)}$ and $\varepsilon Hf_{(t)}$ data, the Palaeoproterozoic crust of the West African Craton displays a dominantly juvenile character (Abouchami, *et al.*, 1990; Block, *et al.*, 2016a; Boher, *et al.*, 1992; Taylor, *et al.*, 1992), with short crustal residence times. Recent studies, however, show a greater Archaean influence in southern Mali and far southeast Ghana (Parra-Avila, *et al.*, 2016; Petersson, *et al.*, 2017; Petersson, *et al.*, 2016). The magmatic processes and geodynamic setting of juvenile crust formation are attributed to a number of scenarios including: an oceanic plateau setting (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992); the accretion of multiple volcanic island arcs (Baratoux, *et al.*, 2011; Dampare, *et al.*, 2008; Pawlig, *et al.*, 2006; Senyah, *et al.*, 2016; Sylvester & Attoh, 1992); the development of a volcanic island arc system within an oceanic plateau setting (Boher, *et al.*, 1992); or intra-cratonic rifting leading to the opening of oceanic domains that were subsequently inverted (Leube, *et al.*, 1990). Alternatively, Begg, *et al.* (2009) suggest that the West African Craton represents the amalgamation of Archaean-aged lower-lithospheric domains.

Geochemical studies reveal the bimodal nature of early volcanism across the craton, with a basal sequence of tholeiitic basalts overlain by calc-alkaline volcanic rocks noted in a number of greenstone belts across the domain (Abouchami, et al., 1990; Baratoux, et al., 2011; Dampare, et al., 2008; Hirdes, et al., 1996; Pouclet, et al., 2006; Senyah, et al., 2016). Lompo (2009) interpret tholeiitic volcanism as a mantle plume event, similar to the differentiated oceanic flood basalt analogy proposed by Abouchami, et al. (1990). Intrusive suites in the southern portion of the West African Craton are frequently described as tonalite-trondhjemite-granodiorite (TTG) series (Doumbia, et al., 1998; Vidal, et al., 2009), based on normative feldspar discrimination plots after Barker (1979). These intrusions have historically been divided into three main groups, as defined by the primary ferro-magnesian mineral and the host domain: 1) variably foliated, amphibole-bearing granitoids hosted in volcanic belts; 2) biotite-bearing granites hosted in meta-sedimentary "basins;" and, 3) potassic, alkaline biotite-muscovite granites and muscovite leucogranites (Baratoux, et al., 2011; Block, et al., 2016a; Egal, et al., 2002; Gueye, et al., 2008; Hirdes, et al., 1992; Leube, et al., 1990; Lompo, 2009; Oberthür, et al., 1998; Tapsoba, et al., 2013; Vegas, et al., 2008). Extensive geochemical characterisation of these intrusions reveal calc-alkaline affinities, negative Nb-Ta, Ti and P anomalies coupled with positive Pb anomalies, with diverse Eu and Sr anomalies and REE fractionation patterns (Baratoux, et al., 2011; Block, et al., 2016a; Egal, et al., 2002; Eglinger, et al., 2017; Feybesse, et al., 2006; Gueye, et al., 2008; Hein, 2010; Leube, et al., 1990; Petersson, et al., 2016; Tapsoba, et al., 2013; Vegas, et al., 2008), bearing striking similarities to classical arc geochemical signatures (Arculus, et al., 1999). TTG magmas and associated lavas are interpreted as the production of partial melting of the down-going slab in a subduction setting (Baratoux, et al., 2011; Peucat, et al., 2005; Pouclet, et al., 2006; Sylvester & Attoh, 1992), with high-pressure, low-HREE TTGs and LILE-enriched diorites in NW Ghana interpreted as the product of delamination of the lower crust and lithospheric mantle (Block, *et al.*, 2016a).

Whilst these lines of evidence suggest that crustal accretion occurred primarily in a subduction setting, a number of characteristics of the craton do not conform with those of modern accretionary orogens. The major site of juvenile crustal growth on the modern Earth is associated with extensive magmatism in magmatic arcs forming along convergent plate margins (Arculus, 1994). Lateral accretion of juvenile terranes or mobile belts against older craton margins results in lateral crystallisation age gradients and isotopic boundaries (Gastil, 1960; Hoffman, 1988; Zeh, et al., 2009). In the West African Craton, significant volumes of juvenile crust form in an oceanic domain, distal to Archaean crustal influences. Furthermore, there are no clear age gradients preserved in sub-parallel greenstone belts emplaced between 2300 and 2180 Ma. The youngest juvenile crust is located in Kedougou-Kéniéba Inlier of Senegal, suggesting a westward gradient (Hirdes & Davis, 2002). Hirdes and Davis (2002) propose a major suture in central Ivory Coast, joining an older eastern sub-province with a 50-100 M.y. younger western sub-province. A ca. 2130 Ma suture between two crustal blocks is hypothesised by Parra-Avila, et al. (2017) for a similar location, based on a comprehensive geochronological transect across Burkina Faso, northern Ghana, southern Mali and eastern Guinea; however, the studies are not in exact agreement, making the extent of the suture somewhat ambiguous. Finally, the metamorphic record and litho-structural architecture do not preserve the same asymmetries, exotic crustal fragments, accretionary prism and metamorphic conditions as modern accretionary systems, thus convoluting the application of a subduction collision model for the West African Craton.

1.3 Synthesis

The greenstone-granite-gneiss terranes of the Palaeoproterozoic West African Craton bear a number of similarities to Archaean provinces. The lithological assemblage, craton-scale shear zones and dominant low- to medium-grade metamorphism are explained by a variety of geodynamic and tectonic models, reminiscent of both Archean and Phanerozoic processes. As such a number of questions arise when attempting to elucidate the geological evolution of the region. These include: What was the geodynamic setting in which the juvenile crust was generated and does it conform to the modern plate tectonic regime? What were the orogenic processes responsible for the amalgamation and stabilisation of such large areas of crust? What are the implications of these finding in understanding the evolution of geodynamic processes and lithospheric properties in the early Earth? In order to answer these questions, the research in this thesis employs a multi-disciplinary, multi-scale approach to understand the crustal and tectonic evolution of the West African Craton during the Palaeoproterozoic, focusing on the Sefwi Greenstone Belt and adjacent domains in southwest Ghana.

1.4 References

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Chapter II

Research aims and thesis outline



2 Research aims and rationale

2.1 Research aims

The ca. 2300–2070 Ma Palaeoproterozoic rocks of the West African Craton are presently underexplored, and poorly represented in the global literature regarding Proterozoic geodynamics. It represents a unique subset of the crustal rocks forme++d after the Archean-Proterozoic transition, with no significant reworking following the craton stabilisation. It thus provides an exceptional window into the geodynamic setting and orogenic processes responsible for juvenile crust formation and preservation and orogenic processes during an enigmatic period of the Earth's history. The Palaeoproterozoic Sefwi Greenstone Belt of SW Ghana in the West African Craton provides the area of focus for this thesis. The primary questions this thesis aims to address include:

1. What was the structural and metamorphic evolution of the Sefwi Greenstone Belt?

2. What were the characteristics of Palaeoproterozoic deformation and metamorphism and what were the possible driving forces? Furthermore, what insight does this provide into the tectonic style of the Eburnean Orogeny and what are the implications for Palaeoproterozoic orogenesis?

3. What was the geodynamic settings which allowed for the formation and preservation of large juvenile crust volumes of the West African Craton, following a hypothesised period of global tectonic quiescence? Do the geodynamic processes responsible for the formation of the West African Craton those active during the Archean or were they formed in a plate tectonic-setting? How do these processes differ from today?

The overarching aim of this thesis is to explore the geodynamic setting of juvenile crust formation as well as the orogenic processes preserved in the West African Craton to gain insight into the evolution of the plate tectonic regime early in the Earth's history. Given the temporally and spatially expansive nature of this aim, this thesis employs a multi-scale and multi-discipline approach combining field evidence, including regional structural and metamorphic mapping, with quantitative *P-T-t* constraints, geochemical and isotopic data. The following thesis outline addresses different aspects of the three primary aims, including the context within existing literature and the methods applied in order to explore different aspects of each question.

2.2 Thesis outline

This thesis is presented as a series of three manuscripts. As each chapter/manuscript is designed as a stand-alone publication, they each require general background information, therefore some repetition and overlap may exist between chapters. Chapter IV was submitted for publication in the Journal of Metamorphic Geology, and was returned for resubmission by May 2018 pending revisions. The manuscripts represented by Chapters III and V are ready for submission at Precambrian Research upon finalisation of the thesis.

2.2.1 Chapter 3: Transtension-related lower crust exhumation in the late stages of the Palaeoproterozoic Eburnean Orogeny, SW Ghana: Evidence for diachronous assembly of the São Luís-West African Craton

A number of difficulties arise when applying a uniformitarian approach to understanding Archean and Proterozoic terranes, whereby their comparison to modern accretionary and collisional orogens contains a multitude of uncertainties due to ambiguous field relationships (e.g. Condie, 1981; de Wit, 1998), the absence or sporadic formation and preservation of plate tectonic indicators (Hamilton, 1998; Stern, 2005) and the influence of the warmer upper mantle temperatures in the early Earth (Gerya, 2014; Sizova, et al., 2014; van Hunen & Allen, 2011). The ca. 2150-2070 Ma Eburnean-Transamazonian Orogeny of the West African-Sao Luis Craton represents the first Palaeoproterozoic orogenic event during a period of global orogenesis between ca. 2100 and 1800 Ma, heralding the first Palaeoproterozoic supercontinent, Nuna (Betts, et al., 2016; Feybesse, et al., 2000; Rogers & Santosh, 2002; Zhao, et al., 2004). A number of conflicting tectonic models are invoked for the West African Craton, with the lack of consensus primarily resulting from extensive overprinting by late transcurrent tectonism, a pervasive greenschist facies metamorphic overprint, poor exposure and limited geochronological constraints. Existing tectonic models include crustal thickening through thrusting and tectonic stacking during subduction-driven lateral shortening (e.g. Feybesse, et al., 2006; Feybesse & Milési, 1994; Ledru, et al., 1994; Liégeois, et al., 1991), archaic dome and basin formation associated with granitoid diapirism and greenstone sagduction (e.g. Lompo, 2009; Lompo, 2010; Pons, et al., 1995), or, alternatively, homogeneous, distributed deformation and metamorphism attributed to a Proterozoic hot-orogen model (e.g. Chardon, et al., 2009; Vidal, et al., 2009).

The primary aim of this chapter was to develop a more detailed regional geological map of the Sefwi

Greenstone Belt and the adjacent Sunyani-Comoé and Kumasi Afema Domains. It presents a synthesis of the geophysical signatures and petrophysical properties of lithologies within the study area, a new regional structural framework and the interpretation of regional airborne geophysical datasets. The integration of field observations, a structural framework and regional geophysical datasets constrained the interpretation of new litho-structural and metamorphic maps of the study area. Metamorphic constraints mentioned in this chapter are presented in Chapter 4, with detailed metamorphic petrography, quantitative metamorphic modelling and U-Pb monazite ages. The tectonic evolution of southern Ghana is discussed in relation to the deformation and metamorphism of granite-greenstone terranes of the West African Craton during the ~2150–2100 Ma Eburnean Orogeny. Finally, structures, strain patterns and distribution of tectono-metamorphic domains are compared to those of Archean and Phanerozoic orogen in order to better understand the characteristics, tectonic setting and style of Palaeoproterozoic orogenesis.

2.2.2 Chapter 4: Petrological and geochronological evidence of collisional orogenesis and lower crust exhumation during the Palaeoproterozoic Eburnean Orogeny, SW Ghana, West African Craton

Metamorphic conditions and gradients within and between different orogenic provinces inform geologists about the tectonic style, crustal architecture and thermal regimes that characterise individual orogens and tectonic settings. The metamorphic record of Phanerozoic subduction-related orogens, including blueschist facies rocks and ultra-high pressure (UHP) metamorphism contrasts the "ordinary" greenschist- to granulite facies metamorphism that dominates most Archean granite-greenstone provinces (Brown, 2006; Brown, 2007, and references therein). Indeed, the major changes in the metamorphic record at the Archean-Proterozoic and the Proterozoic-Phanerozoic transition coupled with changes in structural styles that are interpreted as major changes in the global tectonic regime (e.g. Condie, 1994a; Stern, 2007). Palaeoproterozoic greenstone-granite-graeiss terranes of the West African Craton have often been characterised by apparent homogeneous low-grade metamorphism and transcurrent tectonism attributed to the Palaeoproterozoic Eburnean Orogeny (ca. 2150 – 2070 Ma) (e.g. Junner, 1935; Leube, *et al.*, 1990), with amphibolite to granulite facies metamorphism often interpreted as the product of contact metamorphism (e.g. Caby, *et al.*, 2000; Pons, *et al.*, 1995; Vidal, *et al.*, 1996; Vidal, *et al.*, 2009). Small exposures of orthogneisses and paragneisses are sporadically preserved throughout the craton, with varying interpretations, ranging

from a discordant gneissic basements prompting polycyclic models for the Eburnean Orogeny (e.g. de Kock, *et al.*, 2011; Lemoine, *et al.*, 1990), to the interpretation of gneissic and migmatitic textures as syn-tectonic granitoid emplacement (Gasquet, *et al.*, 2003), with local migmatisation (Caby, *et al.*, 2000). Alternatively, a number of studies have documented transitional greenschist-blueschist facies rocks indicative of cold apparent geothermal gradients (~15°C/km) (Block, *et al.*, 2015; Ganne, *et al.*, 2012), characteristic of modern subduction-related deformation. Proponents of monocyclic orogenesis during the Eburnean Orogeny suggest that low-grade terranes are the supra-crustal equivalents of high-grade rocks, metamorphosed during the same tectonic event and subsequently juxtaposed during structurally controlled exhumation (Block, *et al.*, 2016; Hirdes, *et al.*, 2007).

Existing metamorphic studies of greenstone belts in southern Ghana, including the Sefwi and Ashanti belts, suggest prevailing metamorphic conditions were greenschist to amphibolite facies (Galipp, et al., 2003; John, et al., 1999; Klemd, et al., 2002). Thus far, regional scale upper greenschist to amphibolite facies metamorphism within the Sefwi Belt is poorly elucidated, with elevated metamorphic grades attributed to "the updoming of a large, hidden, underlying potassic pluton" (Hirdes, et al., 1993). This chapter aims to explore the metamorphism and tectonic contacts between tectono-metamorphic domains within the Sefwi Greenstone Belt relative to the adjacent low-grade volcano-sedimentary Sunyani-Comoé and Kumasi-Afema domains. Quantitative assessment of the metamorphic conditions are investigated through the construction of whole-rock pseudosections to constrain *P*-*T* paths in key locations within the study area. New constraints on the timing of metamorphism are provided by new in-situ SHRIMP U-Pb dating of metamorphic monazite. These constraints are used to understand spatial and temporal variations in the metamorphic conditions and relative timing of metamorphism between discrete terranes within the West African Craton. The results of this chapter provide insight into the monocyclic versus polycyclic nature of the Eburnean orogeny, aiding our understanding of the architecture and assembly of the West African Craton. Finally we aim to discuss the significance of these results in understanding Palaeoproterozoic geodynamics.

2.2.3 Chapter 5: Palaeoproterozoic juvenile crust formation in southern Ghana, West Africa: New insights from igneous geochemistry and U-Pb-Hf zircon data

The mechanisms and growth rates of continental crustal genesis have undergone temporal changes since the Hadaean as a product of the Earth's cooling and differentiation. The most important period

of change occurs at the end of the Archean Eon, between 3000 and 2500 Ma, by the end of which more than 50% of the Earth's continental crust had formed (Belousova, et al., 2010; Condie, et al., 2011; Dhuime, et al., 2012; Taylor & McLennan, 1985; Taylor & McLennan, 1995). Archean cratons are systematically characterized by greenstone belts, voluminous juvenile tonalite-trondhjemitegranodiorite granitoids, komatiites and banded-iron formations (Condie, 1994b; Martin, et al., 2005; Moyen, 2011; Sylvester, 1994), commonly bounded by craton-scale shear zones (De Wit & Ashwal, 1997). In contrast, crustal growth in the Phanerozoic is largely limited to accretionary orogens, characterized by calc-alkaline magmatism, lateral age gradients, ophiolite sequences and suturing of juvenile crustal ribbons or fragments onto plate margins (Cawood, et al., 2009, and references therein). The Palaeoproterozoic greenstone belts and granite-gneiss domains of the West African Craton displays both lithological and architectural similarities to Archean provinces, characterised by apparent lack of lateral age gradients and the coeval injection of juvenile TTGs across hundreds of kilometres. The magmatic processes and geodynamic settings for its formation, however, continue to be debated, with authors invoking an ocean plateau setting (Abouchami, et al., 1990; Boher, et al., 1992; Lompo, 2009) or the accretion of volcanic island arcs (Baratoux, et al., 2011; Dampare, et al., 2008; Senyah, et al., 2016; Sylvester & Attoh, 1992).

In this chapter, we investigate the geochemical and isotopic signature of magmatic rocks in the archetypal Sefwi Greenstone Belt. We define the major suites present in the study and use petrological and major and trace element analysis to define possible sources and petrogenetic processes active during crust formation. Geochemical analysis is complemented by zircon U-Pb and Lu-Hf isotopy to constrain the timing of magma emplacement, as well as the timing of magma extraction from the depleted mantle. We present our results within the broader context of magmatism in the West African Craton to discuss the crustal evolution of the craton. Finally, this chapter assesses the viability of various geodynamic settings for crustal growth in the context of the Archean-Proterozoic transition, including subduction-related settings, oceanic plateaus, or tectonically or magmatically over-thickened crust (e.g. Bédard, *et al.*, 2003; Defant & Drummond, 1990; Laurent, *et al.*, 2014; Moyen, 2011; Moyen & Martin, 2012; Smithies & Champion, 2000).

2.2.4 Chapter 6: Discussion and conclusions

This chapter provides a synthesis of the research aims and the significance of the results covered in this thesis. The discussion is composed of a brief summary of literature regarding the West African Craton, including the different geodynamic models proposed for the evolution and preservation of the craton. The significance of the metamorphism and deformation attributed to the Eburnean Orogeny is then considered in the context of the evolution of metamorphism throughout geologic time. We outline the unique nature of orogenic processes recorded in the craton during the Palaeoproterozoic at the commencement of a major period of global orogenesis, heralding the supercontinent era. The subsequent discussion explores the possible geodynamic setting for the formation of juvenile crust preserved in southern Ghana relative to Archean and modern geodynamic processes. Finally we discuss the significance of our findings in the West African Craton within the Palaeoproterozoic rock record in the context of secular cooling of the Earth and the transition to from Archean geodynamics to a modern plate tectonic regime.

2.3 References

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Chapter III

The geology and tectonic evolution of SW Ghana



3.0 Introduction

This chapter focuses on the lithologies and deformation sequence preserved in the Palaeoproterozoic Sefwi Greenstone Belt and adjacent tectono-metamorphic domains of SW Ghana in the West African Craton. We present new litho-structural and metamorphic maps of the study area, which are the product of the integration of field mapping, petrophysical data for individual lithologies and regional airborne geophysical data sets. The multi-scale structural investigation and incorporation of multiple datasets provides insight into the poorly understood structural history, tectonic relationships and processes of a sparsely exposed, yet archetypal greenstone belt of the craton. We consider the deformation style and history in the context of both existing tectonic models for the West African Craton and orogenic models proposed for the early Earth relative to today.

Metamorphic petrology, quantitative metamorphic estimates and metamorphic monazite U-Pb ages referred to in this chapter are discussed in detail in Chapter 4.

Chapter 3: Transtension-related lower crust exhumation in the late stages of the Palaeoproterozoic Eburnean Orogeny, SW Ghana: Evidence for diachronous assembly of the São Luís - West African Craton

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Abstract

This study presents newly interpreted litho-structural and metamorphic maps of the Sefwi Greenstone Belt and volcano-sedimentary domains of southwest Ghana in the Palaeoproterozoic (2300–2070 Ma) West African Craton. The integrated use of field observations, structural mapping and interpretation of regional airborne geophysical data provides new information on the architecture, deformation sequence, overprinting relationships and tectonic evolution of the sparsely exposed region during the Palaeoproterozoic Eburnean Orogeny (2150-2070 Ma). This study reveals tectonically juxtaposed domains of contrasting metamorphic grade comprising high-grade paragneisses and meta-volcanic rocks and low-grade volcano-sedimentary packages, both of which are extensively intruded by multiple generations of granitoids. The earliest deformation is characterised by a ubiquitous, bedding-parallel metamorphic foliation (S₁) preserved throughout the study area, axial planar to poorly preserved isoclinal folds. D₁ NNW-SSE shortening is associated with high-grade metamorphism with partial melting occurring in its later stage. It is interpreted as a crustal thickening event, generating twomica and muscovite granites emplaced between 2092 and 2081 Ma, providing a minimum age for D₁ deformation. D₂ structures are characterized by a moderate to steeply dipping, WSW-ENE to NE-SW striking, penetrative S₂ mineral cleavage. The transtensional nature of the deformation is indicated by the sub-parallel to coaxial nature of F₂ fold axes and L₂ stretching lineations, pitching shallowly ENE-WSW to NE-SW and striking obliquely to the major shear zones. Under a transtensional regime, late-orogenic differential exhumation of the middle and lower crust occurs along NNE-striking extensional detachments, coincident with oblique sinistral displacement along NE-SW regional-scale shear zones. The timing of exhumation is constrained by metamorphic monazite U-Pb ages at ca. 2073 Ma. Overprinting by subsequent E-W shortening is attributed to a change in boundary conditions and is associated with dextral reactivation of regional NE-SW striking shear zones and refolding of earlier structures distal to major shear zones. We propose that the north-western margin of the Sefwi Belt represents the collision of southern Ghana and central and NW Ghana/Ivory Coast which display contrasting tectonic histories. Furthermore, we suggest the accretion of the West African Craton is the product of episodic collisional orogenesis, providing new insights into the evolution of orogenic processes during the Palaeoproterozoic.

Keywords: Palaeoproterozoic collisional orogenesis; transtension; exhumation; West African Craton; Sefwi Greenstone Belt

3.1 Introduction

The applicability of the plate tectonic paradigm and a uniformitarian approach when investigating accretionary and collisional orogenesis in the Precambrian is a continued source of controversy and debate. Modern subduction-related convergent settings are characterised by a number of tectonic and metamorphic features including ophiolites, accretionary mélanges, fold and thrust belts, blueschist facies rocks and ultra-high-pressure metamorphism (UHPM) (Chopin, 2003; Ernst, 2005; Matsuda & Uyeda, 1971; Miyashiro, 1973). The similarities in datasets between the ancient and modern rock record and tectonic processes are interpreted as evidence for the operation of plate tectonics by at least ca. 3000 Ma, based on the appearance of mantle wedge-related geochemical signatures, collages of accretionary-type terranes, palaeomagnetic data indicating ocean crust consumption, metamorphic and structural data indicating collisional orogenesis and numerical modelling of oceanic subduction (Cawood, *et al.*, 2009; Cawood, *et al.*, 2006; Condie, 2008; Condie & Benn, 2013; Pease, *et al.*, 2008; Smithies, *et al.*, 2007). Alternatively, the absence of key tectono-metamorphic features, or differences in their appearance, in Precambrian terranes has prompted a number of authors to propose alternative geodynamic and orogenic processes in the early Earth (e.g. Bédard, *et al.*, 2003; Hamilton, 2011; Stern, 2005; Stern, 2007; van Hunen & van den Berg, 2008).

The continental record of the early Earth is characterised by high crustal growth rates (e.g. Dhuime, *et al.*, 2012), archetypal granite-greenstone terranes (e.g. Condie, 1994; de Wit, *et al.*, 1992) and warmer mantle temperatures (Herzberg, *et al.*, 2010; Korenaga, 2013). Recently, a significant amount of research using numerical, analogue and thermo-mechanical modelling has focused on the role of elevated mantle temperatures and its influence on the mode and viability of subduction and orogenic processes (e.g. Ernst, 2009; Gerya, 2014; O'Neill, *et al.*, 2007; Sizova, *et al.*, 2014; Sizova, *et al.*, 2010). The secular cooling of the mantle corresponds with changes in subduction behaviour, including episodicity and slab break off, as well as changes in lithospheric rheology (Burov & Cloetingh, 2010; Burov, 2011; Rey & Coltice, 2008; Rey & Houseman, 2006). This, in turn, is reflected in the metamorphic record and the strain patterns of ancient orogens (e.g. Bouhallier, *et al.*, 1995; Brown, 2006; Brown, 2007; Choukroune, *et al.*, 1995).

The Palaeoproterozoic rock record represents a unique period of the Earth's history defined by nascent or episodic plate tectonics and ambient mantle temperatures approximately 100 – 200 °C greater than present day (Herzberg, *et al.*, 2010; Korenaga, 2008a; Korenaga, 2008b). Whilst Archaean-like granite-greenstone terranes persist into the Proterozoic, significant changes are observed in the geochemical record at 2500 Ma (Keller & Schoene, 2012; Moyen, 2011) corresponding with the diversification of preserved metamorphic conditions including ultra-high-temperature metamorphism, high-pressure granulite facies metamorphism and rare formation and exhumation of eclogites between 3000 and 2000 Ma (e.g. Brown, 2006; Brown, 2007; Collins, *et al.*, 2004). Elevated magmatism and crustal growth coincides with the development of collisional orogenesis in the Neoarchaean and Palaeoproterozoic, prompting the invocation of a *hot-orogen* model characterised by monotonous HT-LP metamorphic conditions and three-dimensional deformation during the compression of hot lithosphere (Cagnard, *et al.*, 2006; Chardon, *et al.*, 2009; Vidal, *et al.*, 2009). The Palaeoproterozoic (2300–2070 Ma) granite-greenstone terranes of the West African Craton (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992) represents an enigmatic juvenile cratonic domain and a unique opportunity to explore orogenic processes in the early phases of these global transitions.

This study focuses on the tectonic evolution of the Palaeoproterozoic crust of southwest Ghana in the West African Craton, comprising the Sefwi Greenstone Belt and the adjacent Sunyani-Comoé and Kumasi-Afema meta-sedimentary domains. Newly interpreted lithological and structural maps of the region are presented, yielded by the integration of field observations and regional airborne geophysical data. Structural analysis of the region defines a new deformation sequence during the ca. 2100 Ma Eburnean Orogeny, with timing constraints interpreted from existing geochronological data. Field observations of metamorphic assemblages and foliations are presented in a new metamorphic map, which highlights tectonically juxtaposed high-grade domains with low-grade domains. The multiscale, multi-discipline approach provides insight into the tectonic styles and exhumation mechanisms preserved within the craton. Evidence presented in this study for the tectono-metamorphic history of the Palaeoproterozoic Transamazonian-Eburnean Orogeny of the São Luís–West African Craton reveals episodic collisional orogenesis during craton assembly. Finally, we discuss the implications of our findings relative to the Precambrian orogenic record and the secular evolution of metamorphism and thermodynamic properties of the lithosphere in the early Earth.

3.2 Geological Setting

The southern portion of the West African Craton (WAC) comprises the Archaean Kénéma-Man province (3600–2700 Ma) and the juvenile granite-greenstone terranes of the Palaeoproterozoic (2300–1980 Ma) Baoulé-Mossi Domain to the east and north (Bessoles, 1977) (Fig. 3.1a). The



Fig. 3.1: a) The southern portion of the West African Craton showing the study area in the southeast, (modified after Milési, et al., 2004), with inset showing location of Archaean and Palaeoproterozoic domains of the West African Craton, Dark and light green colours represent mafic volcanic rocks and sedimentary rocks intercalated with intermediate to felsic volcanoclastic rocks, respectively. Yellow units comprise unconformably overlying fluvio-deltaic sediments in fault-bounded basins; b) Simplified map from this study showing dominant lithologies, domain names used within the text and major shear zones transecting the SW Ghana. CD: Chiraa Domain; HD: Hwidiem Domain; BGD: Bechem Granitoid Domain; KJD: Kukuom-Juaboso Domain; SWD: Sefwi-Wiawso Domain; SZ: Shear Zone.

Baoulé-Mossi Domain is characterised by elongate volcanic-plutonic greenstone belts and volcanosedimentary basins of the Birimian Supergroup (Fig. 3.1), which are bounded by regional- to cratonscale shear zones (Junner, 1940; Leube, *et al.*, 1990) and voluminous granite-gneiss terranes. Birimian volcanic and volcanoclastic sequences range in composition from basalt to rhyolite, with tholeiitic to calc-alkaline affinities (e.g. Baratoux, *et al.*, 2011; Dampare, *et al.*, 2008; Leube, *et al.*, 1990; Pouclet, *et al.*, 2006). Sedimentary rocks of the Birimian Supergroup include sequences of volcaniclastic greywacke, argillite, shale and chemical sediments, intercalated with intermediate volcanic layers (e.g. Leube, *et al.*, 1990; Oberthür, *et al.*, 1998). Greenstone belts are sporadically overlain by fluvio-deltaic sedimentary sequences of conglomerate, quartzite and phyllite of the Tarkwa Group, which were deposited in narrow, fault bounded basins (Davis, *et al.*, 1994; Eisenlohr, 1989). Multiple generations of Palaeoproterozoic granitoids are gneissic to undeformed, bearing both calc-alkaline affinities as well as geochemical similarities with the tonalite-trondhjemite-granodiorite (TTG) suite, with compositions ranging from dioritic to granitic (e.g. Doumbia, *et al.*, 1998; Gasquet, *et al.*, 2003; Hirdes, *et al.*, 1996).

Prior to the advent of absolute geochronology, the stratigraphic relationship between Birimian metavolcanic and metasedimentary sequences drew a number of conflicting interpretations due to ambiguous, highly tectonised field relationships (e.g. Junner, 1935; Junner, 1940; Milési, *et al.*, 1989; Pouclet, *et al.*, 1996; Tagini, 1971; Whitelaw & Junner, 1929). U-Pb and Pb-Pb zircon crystallisation ages indicate volcanic activity occurred across the craton between ca. 2250 and 2180 Ma (Agyei Duodu, *et al.*, 2009; Delor, *et al.*, 1992; Hirdes & Davis, 1998; Hirdes & Davis, 2002; Hirdes, *et al.*, 1996; Loh, *et al.*, 1999; Oberthür, *et al.*, 1998). Volcanism was considered to be partially contemporaneous with early granitoid emplacement (e.g. de Kock, *et al.*, 2011; Feybesse, *et al.*, 2006; Gueye, *et al.*, 2008; Pawlig, *et al.*, 2006; Tshibubudze, *et al.*, 2013).

The Palaeoproterozoic rocks of Ghana and eastern Ivory Coast include, from south to north, N- to NE-striking volcano-plutonic greenstone belts of the Kibi-Winneba, Ashanti, Sefwi, Bui, Bole-Nangodi and Lawra belts. The intervening metasedimentary domains, often referred in to in the literature as basins, including the Kumasi domains, and the Sunyani-Comoé, or Haute-Comoé domain. A new stratigraphy for the Birimian Supergroup was recently established in southern Ghana by Adadey, *et al.* (2009) and Perrouty, *et al.* (2012), comprising Birimian metavolcanic rocks and subordinate mica-schists of the Sefwi Group, which were emplaced between ca. 2195 and 2170 Ma, and the volcano-sedimentary and sedimentary sequences of the Kumasi Group. Detrital zircon U-Pb ages from the Kumasi Group indicate deposition commenced at ca. 2150 Ma (Adadey, *et al.*, 2009; Davis, *et al.*, 1994; Oberthür, *et al.*, 1998). The Kumasi Group is intercalated with andesitic layers dated at 2142 \pm 24 Ma in the Kumasi Domain (Adadey, *et al.*, 2009), with similar ages indicated for sedimentation in the Sunyani-Comoé Domain (Delor, *et al.*, 1992; Hirdes, *et al.*, 2007). Tarkwaian sediments comprise clastic material sourced from both the Sefwi and Kumasi Groups. The Tarkwaian sediments have a maximum age of deposition, constrained by U-Pb and Pb-Pb detrital zircons of 2132 ± 3 Ma (Davis, *et al.*, 1994; Hirdes & Nunoo, 1994), with the youngest detrital age of 2090 ± 17 Ma (Pigois, *et al.*, 2003). Tarkwaian units in the Ashanti Belt are intruded by granitoids at 2097 ± 2 Ma (Oberthür, *et al.*, 1998). Voluminous amphibole and biotite-bearing, NaO- and CaO-rich TTG and calc-alkaline plutonism occurred between ~2190 and 2130 Ma (Gasquet, *et al.*, 2003; Hirdes, *et al.*, 1996; Tapsoba, *et al.*, 2013). K₂O-rich granitoids and young felsic lavas are emplaced between ~2130 and 2070 Ma in the west of the Baoulé-Mossi Domain in Senegal, along the Archaean-Palaeoproterozoic contact in the Ivory Coast, and to the southeast of the domain in southwest Ghana and southeast Ivory Coast (e.g. Egal, *et al.*, 2002; Hirdes & Davis, 2002; Hirdes, *et al.*, 1992; Petersson, *et al.*, 2016).

Deformation of Palaeoproterozoic terranes of the Baoulé-Mossi Domain developed during the Eburnean Orogeny (Bonhomme, 1962) between ca. 2150 and 1980 Ma (Davis, et al., 1994; Oberthür, et al., 1998). However, there is conjecture about the number and timing of deformation events, and the deformation mechanisms of different terranes (Feybesse, et al., 2006; Lemoine, et al., 1990; Milési, et al., 1989; Vidal, et al., 2009). Throughout the craton, Palaeoproterozoic metamorphism, attributed to the Eburnean Orogeny, is largely restricted to lower-greenschist to amphibolite facies (e.g. Béziat, et al., 2000; Feybesse, et al., 2006; Kříbek, et al., 2008). It has been proposed by some authors that peak metamorphism reached upper amphibolite facies within the contact aureole of plutons (e.g. Pons, et al., 1995). Regional medium-pressure, medium temperature (MP-MT) amphibolite facies metamorphism; however, prevailed in the Sefwi and Ashanti belts (Galipp, et al., 2003; John, et al., 1999), whilst rare, relic high-pressure, low-temperature (HP-LT) assemblages have been documented in NW Ghana (Block, et al., 2015) and eastern Burkina Faso (Ganne, et al., 2012), suggesting a greater diversity in metamorphic conditions than previously assumed. Feybesse et al. (2006) propose that an initial phase of the Eburnean Orogeny corresponds to magmatic accretion followed by partly contemporaneous deposition of flysch-type sequences of the Kumasi Group between 2135 and 2100 Ma. Synchronous Eburnean crustal thickening and medium- to high-grade metamorphism are attributed to thrust tectonism and lateral contractional deformation (Block, et al., 2015; Block, et al., 2016; Feybesse, et al., 2006; Feybesse & Milési, 1994; Ganne, et al., 2012), with subsequent transcurrent tectonism and greenschist facies metamorphism occurring across the craton between 2100 and 2070 Ma (Delor, et al., 1995; Feybesse, et al., 2006; Jessell, et al., 2012; Ledru, et al., 1991; Lompo, 2010).

Juxtaposed terranes with contrasting peak high-grade and low-grade conditions were interpreted to have undergone a concomitant evolution during monocyclic, progressive deformation (Block, *et al.*, 2015; Block, *et al.*, 2016; Galipp, *et al.*, 2003; Hirdes, *et al.*, 2007; Leube, *et al.*, 1990; Loh, *et al.*, 1999), and therefore represented different crustal depths.

Alternatively, some authors propose a two-phase orogenic evolution and consequently divide the orogenic event into the Eburnean I (2200–2150 Ma) and Eburnean II (2110–2088 Ma) deformation events, which were separated by a period of extension and sedimentation (Allibone, *et al.*, 2002a). These events are also referred to as the Eoeburnean and Eburnean, respectively (de Kock, *et al.*, 2012; Perrouty, *et al.*, 2012). They suggest deformation and metamorphism of high- and low-grade terranes is diachronous, with >2150 Ma gneissic crust interpreted as an older basement, indicative of a polycyclic orogenic history (de Kock, *et al.*, 2012; Feybesse, *et al.*, 2006; Hein, 2010; Lemoine, *et al.*, 1990; Tshibubudze, *et al.*, 2013; Vidal, *et al.*, 2009).

The tectonic processes that generate and exhume high-grade terranes are an additional source of discussion. Some authors argue for unit stacking and nappe folding of rheologically strong continental crust during tectonic accretion, reminiscent of modern-style tectonic regimes (e.g. Billa, *et al.*, 1999; Fabre, *et al.*, 1990; Feybesse, *et al.*, 2006; Milési, *et al.*, 1989). Others have argued that increased metamorphic grade is localised to pluton aureoles and found in association with dome and basin strain patterns (Pons, *et al.*, 1995; Vidal, *et al.*, 2009). Alternatively, exhumation of migmatitic rocks is interpreted as the product of buoyancy-driven ascent in a rheologically weak, hot orogenic crust, following homogeneous and distributed crustal thickening and extensive granitoid emplacement (e.g. Caby, *et al.*, 2000; Ganne, *et al.*, 2014; Pouclet, *et al.*, 1996; Vidal, *et al.*, 2009).

3.3 Study area

3.1.1 Litho-structural domains

The geology of southwest Ghana and southeast Ivory Coast comprises the NE-striking Sefwi-Greenstone Belt, which contains a diverse range of metamorphic rocks from greenschist facies to migmatitic packages, flanked to the NW and SE by two vast, low-grade meta-sedimentary rocks of the Sunyani-Comoé and Kumasi-Afema Domains (Fig. 3.1b) (Agyei Duodu, *et al.*, 2009; Feybesse, *et al.*, 2006; Hirdes, *et al.*, 2007). These domains are often referred to as basins in the literature; however, given the pervasive deformation and metamorphism of the meta-sedimentary units and the unknown nature of both the basement and their tectonic contacts, we hereon in refer to them as tectonometamorphic domains. Discrete domains are bounded and intersected by ENE-, NE- or ~N-striking shear zones.

3.3.1.1 High-grade domains of the Sefwi Belt

The Chiraa Domain (CD) is a triangular area in the northeast of the study area, reaching ~30 km in width. It is included as a subdomain of the Sefwi Belt due to metamorphic and structural similarities. The Chiraa Domain comprises highly heterogeneous lithological packages of paragneisses, rare granite orthogneisses and hornblendites that are extensively intruded by foliated two-mica granites and pegmatites. Paragneisses are often observed as elongate rafts and are sometimes migmatitic. They frequently preserve stratigraphic layering with metamorphic assemblages and textures indicating greywackes, pelitic or volcaniclastic protoliths. Metamorphic petrology and quantitative estimates of metamorphic conditions recorded by the domains are detailed in Chapter 4. This contrasts with previous mapping, which classifies the Chiraa Domain as a homogenous granite batholith (Agyei Duodu, et al., 2009; Hirdes, et al., 1993). A biotite-muscovite granite and a muscovite pegmatite in the Chiraa domain yielded zircon U-Pb crystallisation ages of 2093 ± 2 and 2093 ± 2 Ma (Petersson, et al., 2016), respectively, within error of a previous zircon U-Pb age of 2092 ± 2 Ma (Agyei Duodu, et al., 2009) in the north of the domain. The Chiraa Domain is bounded to the west by a poorly exposed, west-dipping NNE-striking shear zone that is rotated into an east-west orientation along its south-dipping southern margin. The eastern margin of the domain is unconformably overlain by Neoproterozoic sediments of the Volta Basin.

The Kukuom-Juaboso Domain (KJD) is a medium- to high-grade domain that forms the core of the Sefwi Belt. This domain is rhomboidal in shape and is 150km in length and 45km at its widest point. The core of the domain is dominated by poly-deformed upper-greenschist to amphibolite facies mafic to intermediate amphibolites and quartz-rich meta-greywackes and metamorphosed volcanosedimentary rocks. It is extensively intruded by multiple generations of granodiorite, granite and maficultramafic plutons. However, very few geochronological dates have been published. Granodiorite intrusions in the north-west of the domain were emplaced at ca. 2135 Ma (Amponsah, 2012). The ENE-striking Hwidiem Shear Zone separates the Kukuom-Juaboso Domain from garnet amphibolites and weakly foliated to gneissic volcaniclastic packages, quartzites and meta-cherts of the Hwidiem Domain (HD). The Hwidiem Shear Zone contains migmatitic mafic orthogneisses characterised by fluid-present melting (detailed in Ch. 4). The Bechem Granitoid Domain (BGD) is dominated by weakly deformed granitoids with compositions varying significantly from quartz diorites to garnetbiotite granites and late potassic biotite granites (Hirdes, *et al.*, 1993). Pyroxenite, gabbro and norite intrusions also form much of the domain displaying a strong spatial associated with high strain shear amphibolite gneiss. Whilst little dating has been undertaken within the domain, a greenschist facies meta-rhyolite flow on the southern margin of the domain yield a U-Pb zircon age of 2189 \pm 1 Ma (Hirdes & Davis, 1998). Southeast of the rhyolite, strongly sheared Tarkwa-like conglomerates with abundant rounded quartzo-feldspathic lithics are observed along the tectonic contact between the Sefwi Belt and the Kumasi-Afema Domain.

3.3.1.2 Low-grade domains of the Sefwi Belt

The NE-SW striking Sefwi-Wiawso volcano-plutonic domain (SWD) represents the southern portion of the Sefwi Belt and extends for ~200 km from Bibiani in southwest Ghana into south-eastern Ivory Coast (Hirdes, *et al.*, 2007; Hirdes, *et al.*, 1993). The domain comprises greenschist facies tholeiitic to intermediate volcanic rocks, including pillow basalts, with minor volcaniclastic and volcano-sedimentary sequences. It is extensively intruded by granites that range in compositions from quartz diorite, monzodiorite, granodiorite and tonalite to biotite granite. These granites were emplaced between 2178 and 2136 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007), pene-contemporaneous with the emplacement of felsic volcanic and volcaniclastic rocks that have ages between ca. 2180 and 2166 Ma (Hirdes, *et al.*, 2007). The domain is bound to the north by the Ketesso high strain zone, to the east by the NNE-SSW-striking Bibiani and Chirano shear zones. The Sefwi-Wiawso Domain is juxtaposed with the Kumasi metasediments at its southern boundary along the Afema high strain zone (Hirdes, *et al.*, 2007). Immature, polymictic conglomerates occur at the easternmost boundary of the domain bound by the Bibiani shear zone (Agyei Duodu, *et al.*, 2009). The detrital material is sourced from both the volcano-plutonic rocks and the adjacent low-grade metasedimentary domains (Hirdes, *et al.*, 1993).

3.3.1.2 Low-grade meta-sedimentary domains/basins

The Sunyani-Comoé and Kumasi-Afema meta-sedimentary domains coincide with the poorest outcrop conditions in the study area. These domains comprise isoclinally folded volcaniclastic rocks, wackes, and argillites, intercalated with felsic to intermediate volcanic and volcaniclastic rocks (Hirdes, *et al.*, 1993). The Sunyani-Comoé Domain has a greater abundance of turbiditic sequences, whilst the Kumasi-Afema Domain is suggested to comprise a more significant volcaniclastic rock component

(Hirdes, *et al.*, 1993). Initiation of sedimentation in the domains is constrained by a 2142 \pm 24 Ma U-Pb zircon crystallisation age of an andesitic layer that is intercalated with metasedimentary rocks in the Kumasi Basin (Adadey, *et al.*, 2009). Metasedimentary rocks of the Kumasi-Afema Domain are intruded by numerous granodiorite and granite plutons between 2136 and 2090 Ma (Adadey, *et al.*, 2009; Hirdes, *et al.*, 1992). Within the Sunyani Basin, elongate, rounded leucogranite intrusions flank the Sefwi Belt to the northwest. Granites and leucogranites emplaced in the Sunyani-Comoé domain parallel to the NW margin of the Sefwi Belt decrease in age from the northeast to southwest. U-Pb zircon crystallization ages of 2088 \pm 1 Ma and 2081 \pm 1 were yielded by the muscovite-bearing Kawtiago granite and the Apouasso granite in southeast Ivory Coast (Delor, *et al.*, 1992; Hirdes, *et al.*, 2007), with corresponding granite emplacement ages of ca. 2093 Ma in the neighbouring Chiraa Domain to the northeast (Agyei Duodu, *et al.*, 2009; Petersson, *et al.*, 2016).

3.3.2 Major Shear Zones

The crustal architecture of the study area is defined by the NE-striking Sefwi shear system (Jessell, et al., 2012), which extends >300 km from central Ghana to southeast Ivory Coast. The southernmost Afema shear zone marks the tectonic contact between the Sefwi-Wiawso volcano-plutonic domain and the Kumasi-Afema meta-sedimentary domain. Delor, et al. (1992) document NWdirected thrust movement along SE-dipping portions of the Afema Shear Zone in southeast Ivory Coast. The Ketesso high-strain zone transects the length of the Sefwi Belt, from central Ghana to south-east Ivory Coast, covered by Neoproterozoic sediments to the northeast and Miocene-Pliocene sediments to the southwest (Hirdes, et al., 2007). It separates the Sefwi-Wiawso and Kukuom-Juaboso domains, forming the southern extent of the Bechem granitoid domain (Hirdes, et al., 2003; Jessell, et al., 2012). Domains of the Sefwi Belt are bounded to the north by the Kenyase-Yamfo shear zone, comprising reactivated sub-vertical and south-dipping shear zones. Within the Kukuom-Juaboso domain, early shear zones are folded around ENE-WSW striking structures, with the Hwidiem shear zone truncating on the Kenyase-Yamfo shear zone. Both the Kenyase-Yamfo and the Ketesso shear zones are transected by discrete, late NNE-striking sinistral faults, the latter being intersected by the extension of the Bibiani shear zone, forming the contact between the Sefwi-Wiawso domain and the Kumasi-Afema meta-sedimentary rocks.

3.4 Methodology and data

3.4.1 Methodology

The new interpreted geological map of the study area integrates traditional mapping techniques with regional geophysical data sets. Outcropping geology was exceedingly sparse due to the presence of an extensive, thick laterite and saprolite sequence (5 to >40m thick) and significant areas of densely vegetated forest reserves covering the study area, precluding continuous mapping. Interpretation of potential field, radiometric and remote sensing data was constrained by lithological, structural and petrophysical information acquired at 484 outcrop localities. This data was complemented by outcrop maps and observations from Hirdes, *et al.* (1993), as well as outcrop databases from mining companies operating in the field area (Kinross and Newmont).

3.4.2 Processing and interpretation of geophysical data

Datasets used included regional airborne magnetic and radiometric data covering SW Ghana were acquired by the Geological Survey Department of Ghana and GoldenStar Resources Pty Ltd between 1984 and 2005 and national Ivory Coast aeromagnetic data collected by Kenting Earth Sciences in 1976. The Ghanaian data is a composite grid of five individual surveys, the details and locations of which are shown in the supplementary information (Fig. A1SI). Four surveys have line spacings of 200–500m and flight altitudes of 70–80 m, depending on the individual survey. Data was collected along E-W flight lines with the data from the south-eastern-most survey (outside mapping area) collected along N-S flight lines. The national airborne magnetic data of Ivory Coast was collected along N-S flight lines that were spaced 500m apart and collected from an elevation of 152 m (Fig. A1SI). Raw data was processed using Geosoft Oasis Montaj software and used for regional interpretations. Processed images are shown in supplementary files (Supplementary Material A; Fig. 2SI–4SI).

3.4.2.1. Aeromagnetic data

Airborne magnetic surveys provide us with a map of magnetic anomalies from which we can derive lithological, structural and kinematic information, as constrained by field observations (Aitken & Betts, 2009; Betts, *et al.*, 2007; Lindsay, *et al.*, 2011; Stewart, *et al.*, 2009). The regional airborne magnetic data covering SW Ghana was gridded as a composite residual magnetic intensity (RMI) grid with a resolution of 100 m. The residual magnetic intensity data covering the Ivory Coast was

gridded to a cell size of 125 m. Conventional reduction of data to the magnetic pole (RTP) becomes unstable near the equator due to the low inclination of the Earth's magnetic field (Li, 2008; MacLeod, *et al.*, 1993), simultaneously generating north-south trending artefacts and inaccurately representing true north-south structures or anomalies (Beard, 2000) (Fig. A1SI). Both residual magnetic intensity datasets covering SW Ghana and the Ivory Coast and the A1 total magnetic intensity (TMI) grid were reduced to the equator (RTE), with a 180-degree phase reversal or inversion (negative RTE) (Fig. ASI3) according to the method of Luo *et al.* (2010). Interpretation of lithology distribution relied on both the negative RTE and analytical signal grids (MacLeod, *et al.*, 1993). Filters included the first vertical derivative, tilt derivative and amplitude normalisation (Automatic Gain Control) aided extraction of structural information and lithological boundaries by highlighting magnetic features and steep gradients in the magnetic dataset (Gunn, *et al.*, 1995; Milligan & Gunn, 1997; Pilkington & Keating, 2009; Verduzco, *et al.*, 2004).

3.4.2.2. Radiometric (gamma-ray) data

Gamma-ray spectrometry data shows the content and distribution of radioactive elements in the rocks and regolith in the uppermost 30cm of the Earth's crust (Minty, *et al.*, 1997; Wilford, *et al.*, 1997). Combined use of topographic information and geochemical data highlights lithology distribution and laterite and regolith cover (Dickson & Scott, 1997; Wilford, *et al.*, 1997), applied during similar mapping in Burkina Faso (Metelka, *et al.*, 2011) and NW Ghana (Block, *et al.*, 2016) and SW Ghana (Perrouty, *et al.*, 2012). U, Th, and K bands were gridded individually at 100 m resolution (Fig. A4SI). The bands are combined in a ternary RBG (Red-Green-Blue) diagram and draped over a shaded image of the digital elevation model (90m resolution; Shuttle Radar Topographic Mission, SRTM, 2000). Band ratio (U/K and Th/K) and Principal Component Analysis (PCA) images were derived from the original bands. Relative abundance ratios were used to distinguish changes in rock type and alteration.

3.4.2.3. Gravimetric data

Gravity data is frequently utilised in 2D and 3D geophysical modelling for constraining the geometry of lithological bodies and structures at depth (e.g. Attoh, 1982; Direen, *et al.*, 2005) Gravimetric data was sourced from the International Gravimetric Bureau (http://bgi.omp.obs-mip.fr/) as EGM08 Free Air and Bouguer anomaly grids. The Bouguer anomaly map (Fig. 2SI) shows long wavelength anomalies is most effective at differentiating regional rock domains, large lithological packages and

the extent of major structures along strike.

3.3.2.4. ALOS PALSAR data

Advanced Land Observing Satellite; Phase Array type L-Band Synthetic Aperture Radar highresolution data (ALOS PALSAR data) collected in 2009 (downloaded from http://www.eorc.jaxa. jp/ALOS/en/) was used as a complementary topographic information dataset to the SRTM. Two polarisation channels (HH + HV) were gridded with a resolution of 25m with a third band ((HH +HV)/2) used in a ternary RGB diagram. Unlike its application in the neighbouring Ashanti belt (Perrouty, *et al.*, 2012), limited structural and detail could be gleaned from this dataset.

3.5 Lithological associations and geological map

The geophysical datasets were integrated within a GIS software for interpretation, according to the method detailed by Metelka, *et al.* (2011). Field observations, including lithological and petrophysical data, were used to constrain lithology distribution, complemented by radiometric and topological maps. Field structural relationships were used to constrain the interpretation of gradients and signals in the processed magnetic data images. Excellent correlation was observed between the structure and topology of lithologies and magnetic fabrics in airborne magnetic data. Integration of field data, petrophysical data and geophysical signatures were used to characterise the principal lithological associations (Table 1) in the new litho-structural map, as outlined below. Measured magnetic susceptibility values for each unit are presented in Appendix B. Gravimetric data facilitated the demarcation of major domains with high anomalies following the greenstone belt and adjacent metasedimentary domains corresponding with moderate to low gravity response, with no visible gradients at the domain contacts. Further interpretation of the gravity data was precluded by the low resolution. Values for magnetic anomalies are derived from the negative RTE image colour bar legend, and, as a product of data processing, are given as negative integers.

3.5.1 Lithologies

3.5.1.1 Basalt/metabasalt/metadolerite

Metabasalts and metadolerites in the field area have been metamorphosed under greenschist facies to amphibolite facies with foliations indicating varying intensity of deformation, often depending on proximity to larger structures such as faults and regional shear zones. Garnet-bearing amphibolites are exposed in the Kukuom-Juaboso domain. Greenschist facies pillow basalts are observed in the Sefwi-Wiawso domain are weakly deformed and pervasively altered, sub-greenschist facies metabasalts. In the airborne magnetic surveys, mafic volcanic rocks correspond with high-amplitude, high-frequency magnetic anomalies, The anomalies are elongate and curvilinear, ranging between 10 and 40 km in length, with bimodal amplitudes of -618 to -509 nT (red hues corresponding to high magnetic anomalies) and -816 to -745 nT(blue hues corresponding to low amplitude), becoming more heterogeneous in proximity to major structures indicating both magnetic fabric, displaying excellent correlation with S₁ foliation measurements suggesting the magnetic data is imaging the earliest foliation. They have a multi-modal magnetic susceptibility range from 0.05 to 13.08–84 x 10⁻³ SI. The dark radiometric signal of basalts indicated depletion in incompatible elements; however, units located along shear zones on domain boundaries show evidence of potassic alteration.

3.5.1.2 Meta-andesite/dacite

Significant areas of andesitic and dacitic rocks occur both within both the Kukuom-Juaboso and Sefwi Wiawso domains of the Sefwi Belt as well as rare intercalated sills in the adjacent metasedimentary domains. Units in the Sefwi-Wiawso domain are variably deformed and metamorphosed under greenschist facies conditions and are often associated with juvenile metasediments. Intermediate lavas are associated with a homogeneous to slightly stippled, low to moderate amplitude magnetic anomalies. They are consistently lower amplitude (-710 to -620 nT) and often longer wavelength in their magnetic response relative to the metabasalts. Late andesitic dykes in the Kukuom-Juaboso domain are observed cross-cutting earlier structures. The meta-andesitic layers have a bimodal magnetic susceptibility of $0.03-0.40 \times 10^{-3}$ SI and $7.41-39.50 \times 10^{-3}$ SI. The radiometric response ranges from low to high intensity, often featuring a magnetic fabric. The units are often associated with distinct U enrichment relative to Th and K in the radiometric data.

3.5.1.3 Rhyolite/felsic volcaniclastic rocks/ meta-rhyolite

Rhyolites and felsic volcanoclastic and pyroclastic units are found in the Sunyani-Comoé and Kumasi-Afema domains as narrow layers, parallel to stratigraphic layering in schists derived from volcanosedimentary rocks. These felsic lava flows and volcaniclastic layers correspond with discrete, elongate bimodal, high- and low- amplitude magnetic anomalies (-783 to -750 nT; -618 to -585 nT), interpreted as parallel to stratigraphic layers and volcanic flows.

table 1. Synthesis subsets of geophys	of the mineratogy, I sical datasets, each	vith 4km scale-bar, u	ttes and geopnysica using the same met	n signatures of the hodology as descril	lithologies represen bed in section 3.4. S	nea on the geologic cale bar is 4km in l	ai map (rig. 5.2). I ength.	
Lithology	Basalt/metabasalt/ metadolerite	Meta-andesite/dacite	Rhyolite/felsic volcaniclasite	Volcaniclastic metasediments	Volcano- sedimentary schist	Composite gneiss	Mafic orthogneiss	Gabbro/norite/pyrox enite complex
Mineralogy	Hbl, qz, (relict Cpx), bt, pl, chl, act,	pl, hbl, qz, ep, ttn, chl, act	qz, chl ,wm,	Qz, pl, chl, wm, ± bt, hbl, grt	qz, chl, ms, gr	qz, bt, pl, wm, grt, ± alsi, tur, gr	pl, hbl, qz, ilm, grt, ep, ttn	hbl (after cpx), pl, chl, ep, act
Magnetic susceptibility range (10 ⁻³ SI)	0.05, 1.01 9.58, 13.08 84.00 (multimodal)	0.03 0.40, 7.41 39.50 (bimodal)	0.02 1.60, 6.25 29.14 (bimodal)	0.06 1.00, 20.01 80.30 (bimodal)	0.02 1.30	0.04 1.30	0.12 9.58 (bimodal)	0.27 2.48
Sample	SB040	SB173	SB044	SB439	SB362	SB398	SB177	SB411
U (ppm) Th (ppm)	1.80	4.60	630	3.70	4.10	4.60	0.40	0.80
K20 (wt%)	1.17	1.54	2.3	2.63	2.79	3.16	0.2	0.81
Airborne magnetic response	Highly magnetic Heterogeneous, elongate near SZ	Low to high intensity	Low to very highly magnetic layers	Low to high, heterogeneous, strong mag. fabric	Low intensity, uniform, smooth	Very low to low intensity, smooth	Low to high intensity, variable	Low to moderate intesity, weak magnetic fabric
Magnetic image			1		24			
Airborne radiometric response	U, Th, K poor (dark), pervasive K alteration along SZ	Enriched in U relative to Th and K	Low U, moderate Th, high K (red - orange)	Low U, K, moderate Th, regolith (light blue)	Low U, Th, mod K, higher K in Kumasi domain (pink - red)	Moderate to high U, Th, K (pink-yellow)	Heterogeneous signal often covered by regolith and	U, Th, K poor (dark)
Radiometric response + shaded DEM								
Map								

Table 1. (continue	ed)							
Lithology	G1: Trondhjemite, tonalite, granodiorite	G2: Trondhjemite and granodiorite	G3a: Granodiorite, quartz diorite	G3b: Potassic granite	G4: Leucogranite	Chemical sediment (Chert)	Polymictic conglomerate	Dolerite dyke
Mineralogy	qz, pl, bt, hbl, ± kfs	qz, pl, bt, hbl, ± kfs, ep	pl, kfs, qz, hbl, bt,	qz, kfs, pl, bt (± hbl)	qz, kfs, pl, bt, ms,	No data	qz, pl, chl, wm, ep, py	ol, opx, cpx, pl, hbl, bt
Magnetic susceptibility	0.15 0.55, 1.01 3.50, 15.47 25.30	0.12 2.03	0.09 1.00, 4.25 - 24.04	0.47 3.88	0.00 0.15	0.35 1.59	0.01 0.50	0.08 0.16, 10.55 87.70
range (10 ⁻³ SI)	(multimodal)		(bimodal)					(bimodal)
Sample	SB252	SB248	SB045	SB023	SB010	SB171	SB417	SB431
U (ppm)	1.90	0.50	2.00	0.60	7.20	no data	no data	0.89
Th (ppm)	4.70	1.50	5.70	0.50	4.20			2.91
K ₂ O (wt%)	1.84	1.24	1.82	4.01	2.51			1.13
Alroome magnetic response	intensity, strong	Low to moderate, homogeneous,	integnity, weak	moderate strength	Low, nomogeneous	moderate intensity	intensity signal	strong, intear signal
Manuatia imaga	magnetic raotiv	SHROOTH	magnetic raviv	magnetic favire				
Magnetic image	(Astro							
Airborne radiometric	K, Th, U poor (dark) heterogeneous	Heterog, relative enrichment in Th;	Low K, high Th, low U (green)	High K, mod U, low Th (red - pink)	High K, moderate Th, U,	Depleted in U, Th and K (dark)	High in K, U, Th (white) and rich in	No clear signal
		o, is poor (units)						
Radiometric + shaded DEM			a costa					
Мар	X							××

Magnetic susceptibilities tend to be low (< $0.60 \times 10^{-3} \text{ SI}$); however, more magnetite-rich felsic volcanoclastic layers have high magnetic susceptibility ($6.25-29.14 \times 10^{-3} \text{ SI}$). Their radiometric signal reveals high K and Th content relative to U.

3.5.1.4 Volcaniclastic/pyroclastic rocks

Immature, volcaniclastic and pyroclastic rocks with subordinate epiclastic meta-greywacke units and rare iron-rich quartzites cover significant proportions of the Kukuom-Juaboso and Hwidiem domains. This unit is metamorphosed at upper greenschist to amphibolite facies conditions. Foliations vary from biotite defined schistose to gneissic, with alternating quartz and biotite/amphibolite bands. Garnet is a common mineral; however, aluminosilicates were not observed. Textural and petrological evidence indicates water-present melting occurred along early shear zones. The magnetic response of the volcaniclastic metasedimentary rocks indicates a heterogeneous distribution of magnetite in alternating highly magnetic and non-magnetic layers, with amplitudes of -585 to -509 nT and -816 to -760 nT, respectively, with short, high-frequency wavelengths. The strong magnetic fabric shows excellent correlation with measured S_0 and S_1 measurements and is therefore used to extrapolate S_1 trajectories. Magnetite-rich strata well define structures within this unit, and excellent correlation was observed between the earliest foliation defined by biotite and/or amphibole and the magnetic fabric displayed in the processed aeromagnetic data. The magnetic susceptibility of this rock was bimodal, with values ranges from low (< 1 x 10^{-3} SI) to extremely high (20.01–80.30 x 10^{-3} SI). The radiometric signal indicates Th enrichment relative to U and K, commonly masked by regolith cover, indicating extensive weathering.

3.5.1.5 Volcano-sedimentary schist

The volcano-sedimentary schist unit comprises greywackes, argillitic schists, graphitic shale and minor intercalated felsic to intermediate volcanic rocks metamorphosed under sub-greenschist to greenschist facies conditions, with elevated metamorphic grade limited to narrow contact aureoles of intruding granitoids. It forms large proportions of the Sunyani-Comoé and Kumasi-Afema domains. Metamorphic assemblages included quartz-chlorite-white mica-feldspar in the Sunyani-Comoé Domain, and quartz-chlorite-actinolite-feldspar meta-greywackes and sericitic argillites in the Kumasi-Afema Domain. Airborne aeromagnetic data covering the metasedimentary rocks of the Kumasi-Afema Domain reveals a smooth low wavelength, low- to moderate amplitude magnetic signal

(-690 to -605 nT), indicating homogeneous distribution of magnetite. Higher magnetic responses and dipole magnetic anomalies are observed proximal to shear zones, associated with localised magnetite formation during deformation. In contrast, the volcano-sedimentary schists and intermediate volcanic rocks in the Sunyani-Comoé Domain are characterised by a prominent bimodal magnetic signal. Smooth, wide and elongate, low-amplitude signals associated with larger exposures of nonmagnetic rocks (-816 to -684 nT) immediately north of Kenyase and then again north of Sunyani, whilst volcano-sedimentary schists along strike of intruding leucogranites often have a high magnetic response characterised by strong ENE-striking magnetic fabrics. These fabrics are characterised by high amplitude (-620 to -545 nT) anomalies with high frequency, which we attribute to the production of magnetite during prograde metamorphic reactions in granitoid aureoles, coupled with structurally controlled fluid flow. Measured magnetic susceptibility values are consistently low (0.02-1.30 x 10⁻³ SI); however, we note that the surface exposures of this unit are often extensively weathered. The radiometric signal reveals K and Th enrichment relative to U in the Kumasi-Afema Domain; however, the airborne radiometric data indicates extensive regolith cover in the Sunyani-Comoé Domain. The volcano-sedimentary schists display pervasive chlorite-sericite-ankerite alteration in proximity to shear zones.

3.5.1.6 Composite gneiss

Large proportions of the Chiraa domain are composed of paragneisses, rare granitic orthogneisses, varying proportions granitic and trondhjemitic melts and pegmatite dykes. The gneissosity in the paragneiss units is defined by alternating mica-rich and quartz-plagioclase-rich bands. Varying modal proportions garnet and micas between layers suggests stratigraphic variability, or possibly bedding, with aluminosilicate minerals limited to rare Al-rich layers. The observed gneissosity in the paragneiss is parallel to the stratigraphic variations, as is the weak gneissosity or biotite foliation in some granite bodies. The magnetic response of the composite gneiss is associated with low-amplitude (-816 to -765 nT), long wavelength anomalies and a weak magnetic fabric that is often truncated by linear responses associated with structural or lithological contacts. Minor proportions of the domain are characterised by a short wavelength, higher magnetic signal (-618 to -550 nT) defining a stronger magnetic fabric, possibly indicating heterogeneous magnetite concentration in stratigraphic layers. Measured magnetic susceptibility values of paragneisses and granites are predominantly low (0.04–1.30 x 10⁻³ SI). The radiometric signal indicates moderate U, Th and K enrichment. A similarly heterogeneous unit is mapped in NW Ghana (Block, *et al.*, 2016), thus the term "composite gneiss" has been adopted for

this study.

3.5.1.7 Mafic orthogneiss

Mafic orthogneisses are high strain units exposed along shear zones marking domain boundaries including along the Hwidiem shear zone and the southern margin of the Chiraa domain. The migmatitic mafic orthogneiss in the Chiraa domain has a metamorphic mineral assemblage of garnet and clinopyroxene with plagioclase-rich leucosomes indicating metamorphism at high-pressure amphibolite–granulite transition. Orthogneisses along the Hwidiem shear zone contain hydrous peritectic minerals, including hornblende, suggesting congruent, fluid-present melting (Weinberg & Hasalová, 2015, and references therein). The magnetic response of the mafic orthogneiss is generally characterised by high-amplitude (-585 to -509 nT), elongated anomalies, often seen subparallel to D_1 and D_2 shear zones; however, amplitudes below -750 nT are associated with migmatitic mafic orthogneisses containing high melting proportions in the western section of the Hwidiem Shear Zone. This reflects magnetic susceptibility of mafic orthogneisses is bimodal with values ranging from 0.12 to 9.58×10^{-3} SI.

3.5.1.8 Gabbro-pyroxenite-norite

This unit contains a variety of lithologies, including variably deformed hornblende-bearing gabbros, norite and clinopyroxene-rich pyroxenite displaying extensive replacement by hornblende, also associated with layered, hornblende-bearing mafic and dioritic intrusive sequences. Pyroxenites located along Hwidiem shear zones display interconnected plagioclase-quartz melt veins. These lithologies are limited to the Kukuom-Juaboso, Bechem and Hwidiem domains in the central Sefwi Belt. These units have a highly irregular magnetic texture with low frequency, low amplitude (-784 to -683 nT) anomalies and a poorly developed magnetic fabric, increasing in strength along faulted margins. They are characterised by low to moderate magnetic susceptibility (0.27 2.48 x 10⁻³ SI) and gamma ray emissions indicate a strong depletion in incompatible elements. Where more strongly deformed, the radiometric signal indicates high U and Th relative to K, indicating extensive lateritic or regolith cover (Dickson & Scott, 1997). This intrusive complex has not been directly dated; however, their chronology relative to the granitoid series is described below.

3.5.1.9 Granitoids

- (1) G1. Trondhjemite, tonalite and rare granodiorite plutons contain biotite as the main ferromagnesian phases, with minor amphibole. They are foliated and locally migmatitic. K-feldspar is present in (micro) granodiorite plutons. G1 granitoids have a low amplitude (-810 to -706 nT), moderate frequency response in the aeromagnetic data and a stippled magnetic texture. Magnetic susceptibility is multi-modal, ranging from 0.15–25.30 x 10⁻³ SI, and are associated with a well-defined magnetic fabric. Their gamma-ray emissions indicate a heterogeneous depletion in radioactive elements.
- (2) G2. These granitoids comprise biotite-bearing, foliated to gneissic trondhjemite and tonalite plutons, with subordinate to rare amphibole. G2 granitoids are rounded to elongate and have a smooth, low to moderate (-783 to -684 nT) magnetic response. Magnetic susceptibility values ranging from 0.12 1.03 x 10⁻³ SI. The relative chronology of G1 and G2 granitoids is deduced from overprinting relationships observed in airborne magnetic surveys, where the magnetic fabric visible within G1 granitoids is truncated along the margins of the smooth, long wavelength, low amplitude responses of G2 granitoids, which suggests relative geochronology of G2 granitoids post-dating G1 granitoids. G2 granitoids in the Bechem Granitoid Domain displayed pervasive silica alteration and magnetite growth proximal to shear zones. Gamma-ray emissions indicate an overall depletion in radioactive elements and relative enrichment in Th. The 50km long Tano River trondhjemitic batholith, located immediately north of Wiawso, is associated with a large low response in the gravity data.
- (3) G3a. Small, irregular elliptical bodies of calk-alkaline, coarse-grained, equigranular amphibolebearing granodiorites and quartz diorites, with subordinate biotite, titanite and K-feldspar, often observed as interlocking bodies. The magnetic signal is sub-elliptical in shape with a heterogeneous, high frequency, low to high amplitude (-717 to -651 nT) magnetic response. This unit has a strong spatial association with G3b granitoids; however, they have distinctly different radiometric signals. Their radiometric signal indicates heterogeneous Th enrichment relative to U and K, with other areas indicating depletion in all incompatible elements. The magnetic susceptibility of these granitoids is bimodal, with values ranging from low (< 1 x 10^{-3} SI), to locally highly magnetic ($4.25 - 24.04 \times 10^{-3}$ SI).
- (4) G3b. These granitoids are porphyritic potassic granites with abundant biotite and K-feldspar. They

contain minor amphibole and accessory titanite. They are characterised as elliptical, sometimes, slightly elongate, with a distinct radiometric signal indicating enrichment in K relative to Th and U. The magnetic response of G3b granitoids is distinguished from G3a granitoids due to its moderate amplitude (-637 to -600 nT) and a more stippled texture, suggesting a heterogeneous distribution of magnetite. G3b granitoid magnetic susceptibility values are low to moderate (0.47–3.88 x 10⁻³ SI) and display a weak to well-defined magnetic fabric in airborne magnetic data.

(5) G4. These leucogranites contain muscovite, +/- biotite, and are often associated with coarsegrained, muscovite-rich pegmatitic bodies. Airborne magnetic data reveals elongate (~ 30 – 40 km), elliptical bodies parallel to the NW-striking tectonic contact between the Sefwi Belt and the Sunyani-Comoé Domain. The magnetic response of G4 granitoids is characterised by smooth, low amplitude (-816 to -750 nT), and long wavelength (>10 km) magnetic responses. Rounded batholiths intruding the Kumasi-Afema Domain are structurally ambiguous in the potential field data. Their magnetic susceptibility values are homogeneously low (< 0.15 x 10⁻³ SI). The radiometric signal for G4 indicates enrichment in K and Th, relative to U. Gamma-ray emissions of large pegmatite bodies indicate enrichment in all incompatible elements. These bodies share similar petrological and petrophysical characteristics to ME3 of Baratoux, *et al.* (2011) and Metelka, *et al.* (2011) and G5 of Block, *et al.* (2016).

3.5.1.10 Chemical sediment (Chert)

Chemical sediment in the field area is largely composed of massive or banded chert, located along the north-western and far western margins of the Kukuom-Juaboso and Hwidiem domains, within the Sefwi Belt. The magnetic response of the chert is associated with asymmetrical, elongate (\sim 5km), low amplitude (-783 to -771 nT), irregular anomalies. Magnetic susceptibility values are generally low (0.35–1.59 x 10⁻³ SI) and the radiometric signal indicates a depletion in all incompatible elements.

3.5.1.11 Polymictic conglomerate and sandstone

Two types of polymictic conglomerate and sandstone units are observed in the field area, confined to the Ketesso Shear Zone and the Bibiani Shear Zone. Where exposed along the Ketesso Shear Zone, this lithology comprises strongly sheared sandstone, siltstone and matrix supported conglomeratic beds with intensely sheared, rounded quartzo-feldspathic clasts. Conglomerates located along the Bibiani Shear Zone contain metasedimentary schist, meta-volcanic and granite-derived lithics, indicating local sources. A steeply dipping S-C fabric with a horizontal mineral stretching lineation is defined by the metamorphic assemblage of either quartz-chlorite-white mica or quartz-chlorite-actinolite, often pervasively altered to sericite and albite. This lithology is associated with narrow, elongate heterogeneous magnetic anomalies, largely controlled by fluid flow along shear zones and has a radiometric signal that indicates high K and Th.

3.5.1.12 Dolerite dyke

Dolerite dykes in the field area are characterised by high-amplitude (-509 nT), linear magnetic anomalies, less than a kilometre in width and 40 to 200 km in length. The dykes crosscut all Palaeoproterozoic formations with three primary orientations at N 340°, N 020° and N 070°. The dipole response characterising some dykes indicates strong magnetic susceptibility and remanence; however, this was not further investigated within this study. In the northeast of the field area, some dyke generations are overlain by Neoproterozoic Volta basin sediments, whilst others intrude the Neoproterozoic sedimentary pile, indicating multiple emplacement events.

3.5.1.13 Volta basin sediments

Covering the north-eastern extent of the field area, the Neoproterozoic Volta Basin sediments unconformably overlie the Palaeoproterozoic units. The sediments comprise horizontal beds of thick cross-bedded sandstone, fine-grained arkoses and rare conglomerates (Junner, 1946; Kalsbeek, *et al.*, 2008). Magnetic anomalies of Palaeoproterozoic unit continue beneath the basin sediments; however, their magnetic signal of is smoothed and slightly reduced in amplitude. The basin margin is most clearly defined in the combined DEM and radiometric image, associated with an elevated topography and extensive K depletion relative to U and Th.

3.5.2 Geophysical response of tectono-metamorphic domains

The study area is divided into a number of tectono-metamorphic domains based on their geology, metamorphic grade and geophysical response. Given the sparsity of outcrop and the rarity of exposure of major shear zones, domain boundaries are interpreted from magnetic lineaments, truncation of magnetic fabrics or contrasting magnetic, radiometric or gravity responses. The tectono-metamorphic domains of the Sefwi Greenstone Belt, including the high-grade Chiraa Domain, correspond with a

NE-striking, high amplitude gravity anomaly (Fig. A2SI). Moderate to low gravity responses flank the belt to the northwest and southeast, corresponding with meta-sedimentary units of Sunyani-Comoé and Kumasi-Afema Domains, respectively. High amplitude gravity anomalies also correspond to the parallel-striking Ashanti Greenstone Belt and the northern Bui Greenstone Belt (Fig. A2SI). The aeromagnetic data provides significantly more detail on the lithology distribution and domain boundaries within the Sefwi Belt based on contrasting magnetic characteristics (Fig. 3.2a).

The north-western margin of the Sefwi Belt is defined by a ~250km long, curvilinear NEstriking magnetic lineament, known as Kenyase-Yamfo Shear Zone, which separates amphibolite facies tectono-metamorphic domains of the belt from low-grade metasedimentary packages of the Sunyani-Comoé Domain. The central, rhomboidal Kukuom-Juaboso Domain is characterised by a strong, E-W to ENE-WSW trending magnetic fabric, predominantly associated with metamorphosed volcaniclastic and greywacke sequences. Magnetic fabrics of the Kukuom-Juaboso Domain are truncated to the west and to the east along NNE-striking structures, interpreted as the domain boundaries. The northern margin of the Kukuom-Juaboso Domain is demarcated by the elongate, narrow, ENE-striking magnetic lineaments of the Hwidiem Shear Zone. Exposures within the shear zone comprise migmatitic meta-basalt and meta-diorites, which are associated with alternative low and moderate-to-high amplitude magnetic anomalies, indicating structurally controlled magnetite destruction and growth during deformation and metamorphism. To the north of the Hwidiem Shear Zone, the Hwidiem Domain is characterised by high, amplitude, high frequency, NE-striking magnetic fabrics intruded by multiple, low magnetic plutons.

In the far northeast of the study area, the Chiraa Domain is characterised by predominantly by a low magnetic signal containing discrete, high magnetic anomalies, consistent with the heterogeneous high-grade paragneiss and granite exposures within the domain. The Bechem Granitoid domain is characterised by a low- to moderate magnetic response, with a weak, heterogeneous magnetic fabric, with high amplitude, linear to curvilinear magnetic anomalies demarcating its southern boundary represented by the ~300 km-long Ketesso Shear Zone. The southern portion of the Sefwi Belts comprises the Sefwi-Wiawso Domain, which is characterised by highly magnetic mafic to intermediate volcanic rocks, featuring a NE- to ENE-striking magnetic fabric and low amplitude, low-frequency magnetic anomalies associated with multiple generations of pluton emplacement.



Fig. 3.2: a) Composite image of the negative reduced to the equator (nRTE) residual magnetic intensity draped over the shaded first vertical derivative (1VD), comprising airborne magnetic surveys from SW Ghana and SW Ivory Coast (survey details in Fig. A1SI); b) Lithological and structural map of the Palaeoproterozoic rocks of SW Ghana. CD: Chiraa Domain; HD: Hwidiem Domain; BGD: Bechem Granitoid Domain; SWD: Sefwi-Wiawso Domain; SZ: Shear Zone. Shear zones are colour coded according to the last deformation event during which they were active. Darker colours of the same lithological unit represent more rocks with a higher magnetic response.

Contrasting all domains within the Sefwi Belt, the Sunyani-Comoé Domain is characterised by alternating high and low NE-striking magnetic domains. Smooth, low amplitude, long wavelength magnetic domains are associated with meta-sedimentary units bounded by highly magnetic lineaments, which are interpreted as shear zones activating during different phases of deformation. Within the high magnetic domains, ENE-striking magnetic fabrics are observed in the metamorphosed volcanosedimentary rocks, surrounding and along strike of elongate, elliptical magnetic lows associated with leucogranitic plutons. Whilst the Kumasi-Afema Domain bearing some lithological similarity to the Sunyani-Comoé Domain, it is characterised by and low- to moderate magnetic response with poorly developed NE-striking magnetic lineaments. The entire study area is cross-cut by late NW-, NNWand N-striking, dolerite dykes associated with narrow, elongate dipole magnetic anomalies.

3.6 Tectono-metamorphic history

3.6.1 Deformation sequence

We distinguish five discrete deformation events in the Sefwi Greenstone Belt and adjacent metasedimentary domains attributed to the Palaeoproterozoic Eburnean Orogeny. The structural framework for these five events is deduced from field observations across the study area, kinematic analysis of processed potential field data and the macroscale geometries of discrete tectonometamorphic domains.

3.6.1.1 D1 NNW-SSE directed shortening

 D_1 is defined by a ubiquitous, penetrative metamorphic foliation present in both high-grade rocks and low-grade rocks throughout the study area. In the Sunyani-Comoé and Kumasi-Afema domains, S_1 trajectories define the axial surface of kilometric NE-SW striking isoclinal folds of thick stratigraphic units, parallel to the crustal-scale shear zones that dominate the architecture of the study area. Conversely, S_1 trajectories of domains within the Sefwi Belt, as well within the Chiraa Domain, are routinely oblique to NE-SW-striking shear zones. In the Kukuom-Juaboso Domain, S_1 trajectories and D_1 shear zones vary in strike between E-W and ENE-WSW, with evidence of overprinting during subsequent deformation. In the high-grade Chiraa Domain, gneissic S_1 trajectories follow an arcuate geometry, elongate toward ~N250, overprinted by subsequent deformation and vertically transposed along domain boundaries.

In the Kukuom-Juaboso Domain (Fig. 3.3a, b, d), S₁ is defined by a shallowly- to steeply-

dipping, schistose to gneissic metamorphic foliation of aligned biotite and amphibole in metavolcanic rocks, para-amphibolites and orthogneisses, which are well imaged in aeromagnetic data. S_1 is parallel to bedding (S_0) in metamorphosed volcaniclastic units. The limited number of isoclinal, centimetric F_1 folds observed in this domain are intrafolial, with fold axes parallel to bedding in metasedimentary units. In the core of the domain, locally migmatitic orthogneisses, interpreted as intermediate metavolcanic sequences, display metamorphic banding dipping moderately (<60°) towards N330 to N350, bearing down-dip amphibole stretching lineations and top-to-the-SSE kinematic indicators (Fig. 3.3a). A rose diagram plotting all S_1 measurements for the Kukuom-Juaboso Domain displays two main trends, striking approximately E-W and N030, respectively (Fig. 3.3d). Amphibolite facies packages and large granitoid bodies of the Sefwi Belt are juxtaposed with the low-grade metasediments of the Sunyani-Comoé domain along the Kenyase-Yamfo Shear Zone; however, D_1 kinematic indicators are largely obscured by later deformation and poor exposure.



Fig. 3.3: (a, b, c) Representative D_1 structures in the study area showing key geometries and overprinting relationships (see Fig. 3.7 for outcrop locations). a) NNW-dipping gneissosity (S_1) in a meta-andesite (SB016) in the Kukuom-Juaboso Domain with foliation-parallel, coarsely crystalline, late- D_1 leucosomes; b) Illustration highlights asymmetry of S_1 -parallel leucosomes with steeply down-dip L_1 amphibole stretching lineations (not shown) indicating reverse top-to-the-
SSE kinematics; c) Bedding-parallel biotite foliation, axial planar to isoclinal F_1 folds in a paragneiss unit (SB359) in the Chiraa Domain, showing low angle relationships with a D_1 shear zone; d) Summary diagram of all S_1 measurements from each domain plotted as a rose diagram to highlight structural trends.

In the high-grade Chiraa Domain, S₁ is defined by a biotite-rich gneissosity parallel to stratigraphic layering in migmatitic garnet-bearing paragneisses (Fig. 3.3c), also observed in subordinate granitic orthogneisses. No reliable L₁ stretching lineations were observed in this domain. S₁ is moderate- to steeply-dipping with a highly variable orientation due to overprinting and transposition (Fig. 3.3d). Migmatitic paragneisses of the domain display stromatic leucosomes, parallel to the S₁ gneissic banding, as well as discordant leucocratic veins and dykes, cross-cutting S₁ foliations. Similarly, migmatitic metabasalts exposed in the Hwidiem Shear Zone display an S₁ gneissosity, defined by aligned hornblende alternative with plagioclase and quartz, also featuring stromatic leucosomes, which suggest partial melting occurred in both domains during, or in the latter stages of, D₁ (Fig. 3.3a). In contrast, in the Sunyani-Comoé and Kumasi-Afema Domains, S₁ is characterised by a chlorite-white mica foliation, parallel to bedding (S_0) in the low-grade metasediments, defining the axial surface of isoclinal F₁ folds. In mafic and intermediate meta-volcanic rocks of the Sefwi-Wiawso Domain, S₁ is predominately characterised by a steeply dipping E-W to NE-SW, chlorite-actinolite-white mica schistosity (Fig. 3.3d) and mineral alignment in biotite and hornblende bearing granitoids. Based on field observations, D₁ is likely characterised as an NNW-SSE directed shortening event associated with isoclinal folding and thrusting, high-grade metamorphism, and partial melting in its latter stages.

 D_1 structures in the airborne magnetic data are primarily preserved within the Kukuom-Juaboso Domain and the Hwidiem Domain, corresponding with high magnetic E-W to ENE-WSW striking lineaments interpreted as D_1 shear zones, often coinciding with a sharp change in the amplitude and texture of the magnetic response. They are parallel to the ENE-striking magnetic fabric of the Kukuom-Juaboso Domain, which corresponds with the bedding-parallel S₁ foliation observed in the field. D_1 shear zones are often refolded around tight ENE-striking folds developed during D_2 , or truncated by later faults. The combined evidence of a bedding-parallel S₁, rare isoclinal intra-folial F_1 folds, coupled with the documented thrust movement along D_1 structures is consistent with lateral contractional deformation and crustal thickening during D_1 NNW-SSE directed shortening. During this event, the Kenyase-Yamfo Shear Zone likely represented an oblique reverse fault; however, due to extensive reactivation, this was not confirmed by field evidence.

The structural expression of D₂ deformation within the study area is highly varied, depending on both the orientation of pre-existing structures and proximity to major D₁ shear zones. In the amphibolite to granulite facies paragneisses of the Chiraa Domain, S₂ is defined by a spaced biotite schistosity, forming the axial surface of tight, gently-inclined F_2 folds (Fig. 3.4a), with fold axes observed approximately parallel to L₂ mineral stretching lineations, defined by elongate kyanite or biotite in favourable lithologies. S₂ progressively rotates in orientation from ~N040 to N270 in an anticlockwise direction, parallel to the arcuate geometries of the S_1 trajectories (Fig. 3.5), dipping steeply to the west in the north of the domain, and shallowing slightly to the south (60–80°S). In the far north of the domain, L_2 and F_2 plunging moderately (18–40°) towards the ~N230, rotating with the metamorphic fabric to an E-W orientation along the southern margin of the domain, plunging shallowly to moderately (~7-44°) towards both the east and west due to subsequent deformation. In a raft of migmatitic mafic orthogneisses (SB177) in the south of the domain, the S₁ gneissosity and plagioclase-rich leucosomes are refolded by centimetric, rootless F_2 folds (Fig. 3.4b), with minor melt accumulation in fold hinges. Megacrystic garnet and clinopyroxene porphyroblasts display S_{1/2} foliation parallel stretching (Fig. 3.4c). Whilst the tectonic contact between migmatitic rocks of the Chiraa Domain and greenschist facies metasedimentary rocks of the Sunyani-Comoé Domain to the west and south is poorly exposed, it is interpreted as significant crustal detachment, juxtaposing domains of contrasting metamorphic grade. Along the western boundary of the Chiraa Domain, the dip direction of the S₂ foliation in both the high and low grade rock progressively rotates counter-clockwise from NW to W and then to the south, along the southern boundary of the domain. The corresponding L2 stretching lineation pitches shallowly to the SW, progressively rotating to the east. The combined structural and architectural features of this tectonic contact is interpreted as a normal, left lateral detachment fault, with foliation rotation attributed to a regional scale drag fold along the larger shear system along strike.

Within the Kukuom-Juaboso and Hwidiem domains, S_1 trajectories and D_1 shear zones are refolded along tight F_2 folds elongate towards N230–N270. S_2 is a moderate- to steeply- (40–82°) NNW to SSE-dipping, discontinuous biotite and/or amphibole foliation (Fig. 3.5), depending on the lithology. The ENE-WSW oriented Hwidiem Shear Zone marks the tectonic contact between the Kukuom-Juaboso and Hwidiem domains. Within the shear zone, metamorphic banding (S_1) and parallel leucosomes in migmatitic mafic orthogneisses are refolded around open, upright F_2 folds, displaying prominent elongation and mineral stretching parallel to F_2 fold axes (Fig. 3.4d). F_2 fold



Fig. 3.4 Representative D_2 structures. a) Tight, SW plunging F_2 folds in a garnet-biotite paragneiss in the Chiraa domain (SB440), refolding pelitic and psammitic layers (S_0) and a bedding-parallel biotite foliation (S_1); b) Migmatitic orthopyroxene-free garnet-clinopyroxene meta-basalt (SB177) on the southern margin of the Chiraa Domain, displaying subparallel relationship between S_2 and transposed, refolded S_1 foliation and plagioclase-rich leucosomes; c) Elongate, megacrystic garnet and clinopyroxene porphyroblasts and plagioclase-rich leucosomes (SB177); d) Open, constrictional F_2 fold refolding S_1 gneissosity in a migmatitic meta-basalt (SB089) in the Hwidiem Shear Zone, displaying a prominent L_2 stretching lineation parallel to the F_2 fold axis; e) L>>S tectonite in the Ketesso Shear Zone displaying pervasive mineral elongation in a quartz dioritic garnet amphibolite; f) Southeast dipping S- C_2 fabrics in the footwall of the Kenyase-Yamfo Shear Zone, associated with steeply east plunge L_2 stretching lineations, indicating normal, left-lateral movement;

g) Prominent, NE-striking sub-vertical S_2 foliation in a pyritic, greenschist facies volcano-sedimentary rocks (SB103) in the Sunyani-Comoé Domain. S_1 and S_2 are orthogonally overprinted by a sub-vertical, N-striking dissolution cleavage (S_3). Inset: h) Sub-vertical NE-striking S_2 crenulation cleavage overprinting a millimetric, flat-lying chlorite-white mica S_1 foliation.

hinges plunge 17–30° towards ~N080, whilst L_2 amphibole stretching lineations plunge near coaxially to N065–N109. This geometry indicates greater hinge parallel elongation relative to hinge-perpendicular shortening, which is characteristic of a transtensional regime (Dewey, *et al.*, 1998).



Fig. 3.5. Summary diagram of D_2 and D_3 structural data from domains within the study area. The first and third columns plot S_2 planes and L_2 and F_2 poles to highlight the coaxial, shallow-plunging relationship of L_2 and F_2 . The second and fourth columns highlight the relatively consistent expression of D_3 shortening throughout all domains.

A similar relationship is observed in a garnet-hornblende paragneiss (SB464) in the Kukuom-Juaboso Domain with isoclinal, near recumbent F_2 folds displaying near coaxial L_2 amphibole stretching lineations. Unlike the Hwidiem Domain, L_2 mineral stretching lineations observed in the core of the Kukuom-Juaboso Domain consistently plunge shallowly (7–34°) towards the west, with more varied orientation proximal to domain boundaries due to later transposition. The coaxial relationship between L_2 and F_2 , their obliquity relative to the major Kenyase-Yamfo Shear Zone attest to constrictional deformation and folding under transtension (Fossen, *et al.*, 2013). The role of constrictional deformation is further demonstrated by the presence of L>>S tectonites in garnetamphibole-bearing dioritic amphibolites (L_2 : 4–11° \rightarrow N230) in the Ketesso high strain zone (Fig. 3.4e). Along strike, metre-scale boudin flow structures are observed in a strongly sheared, highly altered granite (L_2 : 11° \rightarrow N225). According to Sullivan (2013, and references therein), L>>S and L-tectonites are indicative of transtension and ductile deformation in the mid- to lower crust.

 $\mathrm{D_2}$ deformation in the Sefwi-Wiawso, Sunyani-Comoé and Kumasi-Afema domains displays contrasting metamorphic grade and fold style. The NE-striking Kenyase-Yamfo Shear Zone separates the Sunyani-Comoé-Domain and amphibolite facies rocks of the Kukuom-Juaboso and Hwidiem domains. Foliations in the footwall of the shear zone dip moderately to the south (~48-64°) bearing east-pitching mineral stretching lineations (L₂: 52° \rightarrow N092) and normal, left-slip kinematic indicators (Fig. 3.4f), which overprint the S_1 foliation. We, therefore, interpret the shear zone as a D_1 fault reactivated as a left-lateral normal fault under an ENE-WSW tensional regime. In the greenschist facies volcano-sedimentary rocks of the Sunyani-Comoé Domain, bedding and S, are refolded by isoclinal, SW-plunging F₂ folds evident in tight, millimetric crenulations of S₁ by an upright S₂ foliation in F₂ fold hinges (Fig. 3.4 g, h), or locally as a NE-SW oriented, differential $S_{1/2}$ cleavage parallel to the regional scale crustal architecture. In the southern volcano-plutonic Sefwi-Wiawso Domain, D₂ struttres are poorly defined, and are variably expressed as E-W to NE-SW striking, tight folds associated with spaced dissolution crenulation cleavage or a weak chlorite schistosity in favourable lithologies. On the eastern margin of the Sefwi-Wiawso Domain, the NNE-striking brittle-ductile Bibiani Shear Zone separates folded metasediments of the Kumasi-Afema domain in the footwall from polymictic conglomerates in the hanging wall. The Bibiani Shear Zone and the parallel Chirano Shear Zone are interpreted as extensional detachments, bounding a syn-extension graben hosting conglomeratic units containing lithics sourced from both the Sefwi-Wiawso and Kumasi-Afema domains.

The D_2 Hwidiem Shear Zone is characterised by a curvilinear, elongate anomalies with highly varied magnetic amplitudes (-816 to -585 nT) and short wavelengths (500 m – 2 km). The western extent of the shear zone is low magnetic, likely reflecting magnetic destruction during fluid-fluxed partial melting of mafic rocks within the shear zone. Distal to the boundaries of the Kukuom Juaboso Domain, D_1 magnetic fabrics and D_1 shear zones are refolded by regional-scale ENE-striking, tight F_2 folds. The truncation of ~E-W striking high amplitude magnetic fabrics in the domain defines the NNE-striking D_2 shear zone representing the western boundary of the domain. Along this shear

zone, rocks of the Kukuom-Juaboso Domain are juxtaposed with the NE-striking, alternating short wavelength, high amplitude and smooth, low amplitude magnetic signals of volcano-sedimentary schists of the Sunyani-Comoé Domain.

In the north of the field area, alternating NE-striking, high-ampitude and low amplutide magnetic signals of the Sunyani-Comoé Domain are trucated along the boundary of the patchy, weakly magnetic to strongly magnetic lithologies within the Chiraa Domain. Both S_2 foliation trajectories and magnetic anomalies preserved in rocks of the Chiraa Domain progressively rotate counter-clockwise in orientation from NE-striking in the north to E-W striking in the south. This rotation is mirrored by the magnetic fabric of the adjacent greenschist facies sediments of the Sunyani-Comoé Domain. The rotation of D_2 structures is consistent with drag folding associated with sinistral movement along the Kenyase-Yamfo Shear Zone. Additional kinematic indicators of sinistral movement are interpreted from the S-C magnetic fabric in the metamorphosed volcano-sedimentary rocks south of Kawtiego as well as the complex, δ -type shear sense indicated by magnetic units wrapping around a small pluton near Adzopé. In the south of the study area, truncation of early magnetic fabrics occurs along the Bibiani Shear Zone associated with normal movement during D_2 transtension.

3.6.1.3 D3 E-W directed shortening

Early fabrics, folds and shear zones are overprinted by deformation during D_3 , characterised as a progressive E-W directed shortening event in a strike-slip setting. D_3 resulted in distributed accommodation of strain, later transitioning to simple shear-related deformation along narrow shear zones associated with a localised greenschist facies metamorphic overprint. The overprinting relationship between $D_{1/2}$ and D_3 is most clearly evident in internal areas of high-grade domains, with fabrics, shear zones and isograds oriented ~E-W, or approximately parallel to the D_3 shortening axis. In the Chiraa Domain, D_3 is characterised by sub-vertical, N-S striking crenulation cleavage with minor biotite recrystallization along crenulation axes, and open, gently N-plunging F₃ folds (Fig. 3.5).

Microstructural evidence shows crenulation of kyanite-defined L_2 stretching lineations by the overprinting N-S S₃ foliation. The near orthogonal relationship between F_2 and F_3 fold axes has resulted in Type 1 fold interference patterns in both the Chiraa (Fig 3.6a) and Kukuom-Juaboso domains, distal to high-strain zones along domain boundaries. On the NNE-striking western boundary of the Chiraa Domain, earlier fabrics are transposed to a sub-vertical orientation during localised dextral strike-slip movement. The post-crystallisation dextral shearing of garnet-bearing leucosomes in

migmatitic paragneisses (Fig. 3.6b), approximately 5km east of the domain boundary, indicate that D_3 deformation post-dates peak metamorphism. Overprinting relationships in the amphibolite facies, Kukuom-Juaboso domain are demonstrated by the refolding of sinistral drag folds along metre-scale D_2 shear zones by near-orthogonal NNE-striking upright F₃ folds (Fig. 3.6c).



Fig. 3.6: D_3 structures. a) Type-1 fold interference pattern in migmatitic paragneisses in the Chiraa Domain (SB404), illustrating the near orthogonal relationship of ENE-WSW oriented F_2 folds and ~N-S striking S_3 biotite foliations; b) Post-crystallisation dextral shearing of a garnet-bearing leucosome, sub-parallel to S_1 and S_2 foliations in a migmatitic paragneiss on the western margin of the Chiraa Domain (SB147); c) N- to NNE striking vertical S_3 foliation overprinting drag folds generated during D_2 sinistral strike-slip faulting in the southwestern regions of the Kukuom-Juaboso Domain (SB313); d) Prominently elongation of quartz clasts in Tarkwaian-type conglomerates and chlorite-white mica $S-C_3$ fabrics (SB418) indicating dextral shearing along the Ketesso Shear Zone.

In the low-grade metavolcanic and metasedimentary domains, D_3 deformation generates a N-S to NE-SW oriented dissolution cleavage, most poorly developed in the Sefwi-Wiawso Domain. D_3 E-W shortening transposes tectonic fabrics proximal to domain boundaries to sub-vertical orientations, reactivating pre-existing shear zone with favourable orientations. In the high-strain Ketesso-Shear Zone and the Kenyase-Yamfo Shear Zone chlorite-muscovite NE-SW S-C₃ fabrics, sigmoidal asymmetries and sub-horizontal stretching lineations indicate dextral shearing under greenschist facies conditions. Conglomerate units within the Ketesso Shear Zone display prominent shearing of clasts coupled with the development of chlorite-white-mica S-C₃ fabrics (Fig. 3.6d). In contrast, the NNE-striking Bibiani on the eastern boundary of the Sefwi Wiawso Domain is at high angles to D₃ shortening and is, therefore, less favourable for strike-slip reactivation. Instead, Tarkwa-type conglomerates deposited during D₂ transtension were thrust over low-grade metasediments of the Kumasi-Afema Domain during D₃.We correlate D₃ deformation with brittle-ductile shearing and west-side-up movement associated with greenschist facies metamorphism documented in the Chirano lode horizon, parallel to the Bibiani shear zone (Allibone, *et al.*, 2004).

 D_3 deformation in the airborne magnetic data sets is associated with high amplitude, elongate NE-striking anomalies between 250 and 300 km in length, associated with the linear Ketesso Shear Zone and the curvilinear Kenyase-Yamfo and Afema Shear Zones. These anomalies truncate those associated with the Hwidiem Shear Zone and the southern margin of the Chiraa Domain. Gentle N-S F_3 folds reform early magnetic fabrics within the Kukuom-Juaboso, Hwidiem and Chiraa Domains.

3.6.1.4 D4: N-S shortening - Brittle ductile deformation and NNE-SSW sinistral shear zones

 D_4 is associated with NW-SE-directed shortening associated with brittle-ductile deformation. This deformation event did not significantly influence the macro-scale crustal architecture of the study area. D_4 generated late sinistral oblique reactivation of the NNE-SSW Bibiani and Chirano shear zones, deflected existing fabric and faults in the Kumasi-Afema Domain. Pseudotachylites in the northern areas of the Bibiani Shear Zone crosscut an earlier east-west mineral foliation in a meta dacite (SB051). Immediately west of the Chiraa Domain, NE-SW striking structures are truncated by a narrow N-S sinistral shear zone, with C_4 -S fabrics developed in a graphitic schist. In the aeromagnetic data, NNS striking D_4 faults, including the reactivated Bibiani Shear Zone, truncate the Ketesso Shear Zone immediately to the north, whilst a smaller N-S shear zone west of Chiraa truncates reactivated D3 shear zones in the Sunyani-Comoé Domain.

3.6.1.5 D5 WNW-ESE shortening; Reidel shear zones

Late NW-SE, NE-SW and E-W oriented structures, which form conjugate fault sets developed during D5 WNW-ESE shortening. Dextral displacement commonly observed along E-W striking shear zones and sinistral movement long NW-SE striking shear zones. Similar late-stage deformation was documented in neighbouring eastern Ivory Coast (Delor, *et al.*, 1992; Hirdes, *et al.*, 2007; Lüdtke, *et al.*, 1998). The magnetic expression of D_5 faults is interpreted from the truncation

and lateral displacement of pre-existing magnetic lineaments and fabrics. Faults are often localised, with a maximum length of 10 km.

3.6.2 Metamorphic history

We present a new metamorphic map for the study area (Fig. 3.7) based on field observations of petrological assemblages and metamorphic textures. Here we describe the key assemblages and metamorphic relationships for each domain (mineral abbreviations after Whitney & Evans, 2010). More detailed petrological analysis and quantitative *P*-*T* estimates and geochronological constraints from in situ monazite SHRIMP U-Pb ages are presented in Chapter 4.

Rocks from the Chiraa and Kukuom-Juaboso domains preserve high-grade metamorphic assemblages. The peak pressure assemblages of paragneiss sample SB398 comprises Bt + Grt + Ky + Pl + Ph + Rt + Qz, whilst garnet amphibolites comprise Hbl + Grt + Pl + Ilm + Qz indicative of high-pressure amphibolite facies metamorphic conditions. Initial conditions of 10.0–11.5 kbar, 580– 650 °C, deduced from unzoned garnet porphyroblast compositions are associated with low apparent geothermal gradients (15–17 °C/km) (detailed in Chapter 4). Migmatitic paragneisses of the Chiraa Domain contain leucosomes parallel to the S₁ foliation, often refolded by F2 folds or feeding into dykes that cross –cut both D1 and D2 structures. This suggests that partial melting was late-D₁ to syn-D₂. Anatexis in the Chiraa Domain matches a higher apparent geothermal gradient of 20–25 °C/ km, consistent with the peak metamorphic assemblage observed in a migmatitic metabasic rock in the same domain of Pl + Grt + Cpx + Hbl + Ilm + L, which suggests anatexis occured at >10 kbar and >700 °C (Pattison, 2003). Conversely, garnet amphibolites from the Kukuom-Juaboso Domain do not undergo partial melting. In contrast, metamorphosed volcano-sedimentary rocks of the Sunyani-Comoé Domain preserve a peak assemblage of fine-grained chlorite-white mica \pm actinolite.

Field observations revealed a localised greenschist facies metamorphic overprint along the boundaries of the amphibolite facies and migmatitic domains and along major shear zones, whilst little to no retrogression of high-grade assemblages was observed away from domain boundaries. This suggests exhumation and juxtaposition of domains of contrasting metamorphic grade occurred prior to the greenschist facies metamorphic overprint. In the Kukuom-Juaboso Domain, retrogression the garnet amphibolite sample (SB344) is marked by the localised chlorite growth along the foliation planes and on the rim of garnets and the overgrowth of titanite around foliation-parallel ilmenite (Chapter 4). The incomplete replacement of hornblende and garnet by chlorite and syn-kinematic growth of geophysical datasets. and juxtaposition of domains of contrasting metamorphic grade during D_2 are highlighted. S_3 foliation trajectories are extrapolated from both field observations and interpretation of display most recent fault movement, with the exception of the western boundary of the Chiraa Domain and the central segment of the Kenyase-Yamfo Shear Zone where normal movement Fig. 3.7: Structural and metamorphic map of the study area displaying S₁ foliation trajectories, major shear zones and the locations of samples mentioned in the text. Major shear zones



chlorite on foliation plane indicates late deformation is associated with a clockwise P-T path from upper amphibolite facies conditions to upper greenschist facies conditions (Spear, 1993). This is consistent with the clockwise P-T paths displaying simultaneous cooling and decompression and an overprinting transcurrent-related, retrograde mineral assemblage of chlorite-carbonate-epidote-albite, documented in the Kukuom-Juaboso Domain by Galipp et al (2000; 2003).

The timing of the initiation of cooling and exhumation in the high-grade paragneisses (samples SB398 and SB389) is constrained by *in situ* monazite U-Pb SHRIMP ages of 2073 ± 2 Ma and 2074 ± 3 Ma (see Chapter 4). Monazite grains were located primarily in the matrix, elongate parallel to the main tectonic fabric, or in the case of SB398, parallel to the L₂ kyanite stretching lineation. The monazite age is interpreted as the timing of exhumation initiation, post-dating zircon U-Pb crystallisation ages of 2093 ± 2 Ma and 2092 ± 2 Ma yielded by granites in the Chiraa Domain (Petersson, *et al.*, 2016). This relationship is consistent with observations from many high-grade terranes in which monazite populations are younger than zircon populations from the same area (Kelsey, *et al.*, 2008).

3.6.3 Structural-metamorphic map

Figures 3.7 and 3.8 display the trajectories of the S_1 foliations over the new tectono-metamorphic map of the study area and illustrates the distribution of metamorphic facies, including localised water-fluxed incongruent partial melting along the Hwidiem shear zone within the Sefwi Belt. This methodology is utilised by Block, *et al.* (2016) for Palaeoproterozoic rocks exposed in NW Ghana and is applied here to facilitate comparison of strain patterns in comparable terranes in the West African Craton. Fault generations are colour-coded according to their primary deformation phase or the event during which they were last reactivated. This map illustrates that:

- The early S_1 and subparallel S_2 fabrics preserve an E-W to ENE-WSW strike within the higher grade subdomains of the Sefwi Belt, complemented by the major trends displayed in rose diagrams of S_1 measurements for the Kukuom-Juaboso and Hwidiem Domains in Figure 3.4. The Kenyase-Yamfo Shear Zone along the north-western margin of the belt demarcates the transition from ENE-WSW striking foliation trajectories to the predominantly NE-striking S_1 trajectories of the Sunyani-Comoé Domain. The shear zone is likely to have been inherited from either the primary crustal architecture or D_1 and has been reactivated and transposed during subsequent deformation. Within the Sefwi Belt, smaller D_1 shear zones and isograds associated with partial melting at the upper amphibolite to granulite facies conditions commonly strike



Fig. 3.8. (a) Detailed structural-metamorphic maps of the Chiraa Domain and (b) the intersection of the Hwidiem Shear Zone and the Kenyase Shear Zone (see Fig. 3.6 for locations). Juxtaposition of amphibolite facies and migmatitic rocks with greenschist facies volcano-sedimentary sequences occurs along normal oblique D2 faults, which are reactivated and transposed during D3 deformation.

ENE and are variably refolded or truncated by younger structures. The NNE-striking western margins of the Chiraa and Kukuom-Juaboso domains coincide with the most significant lateral metamorphic breaks, which are attributed to D_2 ENE-WSW transtension (Figs. 3.7 and 3.8).

- Refolded S₁ trajectories with the Kukuom-Juaboso Domain reveal regional-scale tight F_2 folds, generating a low-angle relationship between S₁ and S₂. D₂ structural measurements (Fig. 3.5) indicate a near coaxial relationship of shallowly E- to NE- and W- to SW- plunging L₂ and F₂ measurements, most prominent in the higher metamorphic grade domains. Figure 3.8 highlights the obliquity between the E-W to ENE-WSW striking S_{1/2} measurements in the Chiraa Domain (Fig. 3.8a) and the Kukuom Juaboso Domain (Fig. 3.8b) and the NE-SW strike of the major shear zones. The obliquity of this relationship is characteristic of terranes deformed under a transtensional regime.
- Sheared S₁ foliation trajectories form S-C fabrics which indicate sinistral movement along multiple D₂ shear zones in the Sunyani-Comoé Domain, often occurring along strike of elongated leucogranitic plutons between the townships of Chiraa and Kawtiego and toward the Ghanaian-Ivorian border (Fig. 3.7).
- Rocks of the Kukuom-Juaboso domain are transected by and entrained in the Ketesso highstrain zone during D₃. Upper amphibolite rocks are documented in the extension of the high strain zone on the Ivorian side. Hirdes, *et al.* (2007) suggest that the rocks represent a highgrade, deeper crustal equivalent of the Sunyani-Comoé domain metasediments whilst we contended that they form part of the Kukuom-Juaboso domain. E-W shortening during D₃ refolded S_{1/2} foliation trajectories within the Kukuom-Juaboso Domain whilst transposing its NNE-striking western boundary to a sub-vertical orientation. Plots of S₃ measurements distal to shear zones indicate a relatively consistent N-S strike attributed to pure shear deformation, whilst simple shear-related deformation localised along major NE-SW-striking shear zones generating dextral S-C₃ fabrics (Fig. 3.5).
- Foliation trajectories on either side of the NNE-striking Bibiani Shear Zone display drag folds associated with sinistral shearing along the fault. Truncation of the Afema Shear Zone at the southern extent of the Bibiani Shear Zone suggests the latest reactivation post-dates D₃ shearing along the Afema Shear Zone.

3.7 Discussion

3.7.1 Timing of Eburnean deformation in the Sefwi Belt

Palaeoproterozoic crust exposed in southwest Ghana comprises tholeiitic to calc-alkaline volcanic suites and TTG-like to calc-alkaline plutons suites was emplaced between 2190 and 2130 Ma, recognised as a phase of magmatic accretion (e.g. Amponsah, 2012; Hirdes & Davis, 1998; Hirdes, et al., 2007; Hirdes, et al., 1993; Loh, et al., 1999; Taylor, et al., 1992). Subsequent deformation and metamorphism attributed to the Eburnean Orogeny are bracketed across the craton between ca. 2150 and ca. 1980 Ma (e.g. Davis, et al., 1994; Feybesse, et al., 2006). Deposition of metasediments in the Kumasi-Afema Domain, considered to be partially contemporaneous with sedimentation in the Sunyani-Comoé Domain, is constrained by the youngest detrital zircon age of 2154 ± 2 Ma (Oberthür, et al., 1998), with intercalated andesitic flows dated at 2142 ± 2 Ma (Adadey, et al., 2009), and latesedimentary emplacement of granodiorite plutons at 2136 ± 19Ma (Adadey, et al., 2009). Perrouty, et al. (2012) bracket Kumasi Group deposition between 2154-2125 Ma, in agreement with a period of tectonic extension and emplacement of alkaline granites and quartz monzonites in the Sefwi-Wiawso domain between 2150 and 2130 Ma proposed by Hirdes, et al. (2007). Figure 3.9 illustrates the interpreted tectonic setting and relative distribution of key terranes in southwest Ghana prior to the commencement of D₁ deformation at ca. 2100 Ma. The tectonic contacts between domains of the Sefwi Belt and the adjacent metasedimentary domains have been significantly reworked during subsequent deformation, and therefore their nature remains unclear. Gravity data; however, defines major litho-tectonic domains, with high gravity anomalies matching the boundaries of the volcanoplutonic and high-grade domains of both the Sefwi and Ashanti belts, and low to moderate strength anomalies associated with the metasedimentary domains, likely reflecting the architecture generated during early crust formation.

In the study area, D_1 deformation is characterised by a ubiquitous bedding parallel foliation, development of E-W oriented regional scale thrusts and kilometric folds in association with HP-MT temperature metamorphism (Fig. 3.9b), consistent with crustal thickening (England & Thompson, 1984). These observations are consistent with field observations of a subparallel relationship bedding and S₁ in the Sunyani-Comoé Domain exposed in southeast Ivory Coast (Delor, *et al.*, 1992; Delor, *et al.*, 1992b). The timing of D₁ crustal thickening remains unconstrained in this study; however, *P-T-t* paths indicate rapid burial of supracrustal rocks precedes anatexis in the Chiraa Domain. We,



Fig. 3.9. a) Block diagram illustrating the interpreted tectonic setting of greenstone belts prior to D_p, highlighting the close spatial relationship between the grey shear zones represent NE-SW striking structures transposed or reactivated during subsequent D_3 E-W shortening. D_3 deformation was progressively resulting in collision along the NW margin of the Sefwi belt, associated with folding and low angle thrust faults, crustal thickening and subsequent lower Domain; d (p. 95)) Metamorphic block diagram highlighting the differential exhumation of high-grade rocks of the Chiraa and Kukuom-Juaboso domains relative to greenschist facies rocks of the Sunyani-Comoé Domain along west dipping detachments and NE-striking sinistral oblique shear zones. Projected Sefwi and Ashanti belts and an unknown distance between the Sunyani-Comoé Domain and the Sefwi Belt prior to D, collision; b) D, NNW-SSE shortening crustal anatexis; c (p. 95)) Initiation of D, ENE-WSW sinistral transtension. Figure displays the distribution of major lithological packages following generated during D_1 anatexis. D_2 transtension results in the formation of grabens at high angles to major shear zones, or as narrow dilational jogs within within the Kukuom-Juaboso Domain display a coaxial relationship pitching shallowly to the WSW or to the ENE in the Hwidiem Shear Zone and Hwidiem the emplacement of leucogranites (ca. $290\overline{2}$ - 2081Ma) parallel to the NW margin of the Sefwi Belt within the Chiraa and the Sunyani-Comoé Domains, shear zones, hosting conglomeratic Tarkwa-like units (yellow unit). Constrictional F, fold axis (grey trajectories) and L, stretching lineations (black arrows) localised into these shear zones and was associated with a greenschist facies retrograde metamorphic overprint.



therefore, suggest that the minimum timing of D_1 is at ca. 2093 Ma indicated by the oldest granite crystallisation age from the Chiraa domain (Petersson, et al., 2016). D₁ field observations potentially correspond with the shear-related, bedding-parallel foliation documented in the neighbouring Ashanti belt by (Allibone, et al., 2002a; Allibone, et al., 2002b). The timing of early Eburnean deformation is poorly constrained in southern Ghana, with authors suggesting it extended from 2187 to 2158 Ma (D1; Perrouty, et al., 2012) or 2116 to 2088 Ma (Eburnean II; Allibone, et al., 2002a). It must be noted that Allibone, et al. (2002a) suggest that metamorphism of Birimian mafic volcanic rocks was attributed to extensive magmatism between ca. 2200 and 2150 Ma, referred to as Eburnean I. Alternatively, thrustrelated crustal thickening in southern Ghana is hypothesised by Feybesse, et al. (2006) between 2130 and 2100 Ma. This likely corresponds with monazite and vein-hosted zircon growth at 2102 ± 1 Ma and 2104 ± 3 Ma, respectively, in migmatitic orthogneisses in the southern extent of the Ashanti belt (Loh, et al., 1999). Lobjoit (1964) suggested that the deflection of the strike of meta-sedimentary rocks around a foliated tonalite pluton east of Bibiani was the product of "forceful emplacement" of the pluton. We instead suggest that the structural evidence may indicate syn- to late-kinematic emplacement of the pluton during D₁ deformation, providing a minimum age for the commencement of D₁ deformation of ca. 2116 Ma (Hirdes, *et al.*, 1992).

The transition from a compressional to a tensional regime and transcurrent D_2 tectonism is constrained by monazite crystallisation ages hosted in both L_2 kyanite porphyroblasts and the matrix of a biotite-sillimanite magmatic gneiss, which both yielded SHRIMP U-Pb ages of ca. 2073 Ma. This age records the initial stages of cooling along the retrograde path and the initiation of ENE-WSW transtension-related exhumation during D_2 (Fig. 3.9c,d). This is consistent with the findings of Jessell, *et al.* (2012) who concluded that transcurrent tectonism along the NW margin of the Sefwi Belt in the Sunyani-Comoé domain post-dates the emplacement of the Apouasso leucogranite at ca. 2081 Ma. Whilst Jessell, *et al.* (2012) suggest that the elongation of the leucogranites occurred during dextral shearing, the combined evidence of constriction related structures and the macro-geometries of juxtaposed metamorphic domains (Fig. 3.9d), we conclude that D_2 deformation drove oblique sinistral reactivation of major D_1 tectonic boundaries, in agreement with sinistral D_2 deformation interpreted by Feybesse, *et al.* (2006),and sinistral kinematic indicators observed along strike in southeast Ivory Coast (Vidal, *et al.*, 2009). Furthermore, we propose that the conglomerate and sandstone sequences west of the Bibiani Shear Zone were deposited in NNE-striking half-graben structures that developed during D_2 transtension. Younger D_3 deformation was associated greenschist facies metamorphism and dextral reactivation of the Sefwi shear system during E-W shortening (Grey shear zones: Fig. 3.9d). The timing of D_3 remains unconstrained; however, it is potentially correlated with retrograde greenschist facies metamorphism and hydrothermal xenotime crystallisation at 2063 ± 9 Ma post-dating sinistral shearing in the Ashanti Belt (Pigois, *et al.*, 2003).

3.7.2 Rheology, tectonic style and exhumation

Strain patterns and lateral changes in metamorphic grade are widely investigated in Precambrian provinces due to the insight they provide into the rheological structure of the lithosphere and its influence on tectonic style and exhumation processes during orogenesis. In Phanerozoic orogenic belts, lateral tectonic forces are accommodated through thrust-related, heterogeneous crustal thickening and unit stacking. Conversely, Archaean and some Palaeoproterozoic terranes are associated with warmer lithospheres and elevated continental geotherms, which results in weaker and more buoyant lithosphere (Griffin, et al., 1998). Under such conditions, convergence is accommodated through homogeneous crustal thickening, limited topography and lateral ductile flow (Cagnard, et al., 2006; Cruden, et al., 2006; Rey & Houseman, 2006). Alternatively, crustal thickening is attributed to the diapiric emplacement of voluminous buoyant granitoids and the sagduction of negatively buoyant greenstone terranes, resulting in the characteristic dome and basin strain pattern, with subsequent lateral shortening accommodated by craton-scale strike-slip shear zones (Bouhallier, et al., 1993; Choukroune, et al., 1995). In the Sefwi Greenstone Belt, we argue that the combined evidence of a bedding-parallel foliation (S₁), early D₁ thrust faulting and initial HP-MT metamorphism indicates burial of supracrustal rocks reflects the early stages of modern collisional orogens (England & Thompson, 1984). During D₁ deformation, we propose that large-scale thrusting leads to heterogeneous crustal thickening and burial of supracrustal rocks during horizontal convergence, early in the tectonic history of SW Ghana. This is consistent with hypothesised accommodation of crustal thickening by nappe stacking (Allibone, et al., 2002b; Feybesse, et al., 2006; Milési, et al., 1992). Such evidence does not conform to dome and basin geometries, homogeneous crustal thickening and diapirism-related deformation as proposed for areas of the West African Craton (e.g. Delor, et al., 1995; Lompo, 2009; Vidal, et al., 1996; Vidal, et al., 2009).

HP-LT and HP-MT metamorphic conditions and low geothermal gradients are rare in ancient orogenic domains (Brown, 2007). Whilst few in number, tectono-metamorphic studies of

Palaeoproterozoic rocks in the West African Craton sporadically preserve peak pressures of 10 - 14 kbars associated with relatively cool apparent geothermal gradients, indicating a minimum crustal thickening of 33 - 50 km, recorded in NW Ghana (Block, *et al.*, 2015), within the Sefwi Belt (Ch. 4, this study), eastern Burkina Faso (Ganne, *et al.*, 2012) and along the tectonic contact between the Archaean Kénéma-Man Domain and the Palaeoproterozoic Baoulé-Mossi Domain (Pitra, *et al.*, 2010). Documentation of such conditions in the West African Craton suggests that localised areas of the lithosphere were cool enough and strong enough to support moderate amounts of crustal thickening, suggesting the development of thermo-mechanical properties similar to that of the Phanerozoic lithosphere (Rey & Houseman, 2006).

The metamorphic record of the study area reveals high-pressure mineral assemblages including two garnet amphibolites documented in the Kukuom-Juaboso Domain, a Grt-Ky-Rt bearing paragneiss (SB398) and an orthopyroxene-free garnet-clinopyroxene migmatitic mafic orthogneiss (SB177) hosted within the Chiraa Domain. The phase assemblage of the mafic orthogneiss is consistent with metamorphism at pressures greater than 10 kbar and 700 °C (Pattison, 2003). It was noted by Pattison (2003), however, that whilst this phase assemblage is characteristic of high-pressure granulite facies metamorphism, it was also possible to generate similar assemblages under amphibolite facies conditions. Metamorphic constraints coupled with field observations, including stromatic leucosomes feeding into discordant veins and dykes cross-cutting S₁ gneissosity indicate partial melting occurred at mid- to lower crustal depths towards the end of D_1 deformation and the transition to D_2 . The complex foliation and lineation structural pattern associated with D₂ deformation indicates a switch in the dominant strain axes and a change in the tectonic mode. Field observations from throughout high-grade domains of the Sefwi Belt are indicative of a transtensional regime. These include shallowpitching, coaxial F₂ folds and L₂ stretching lineations oriented ~ENE-WSW, indicative of transtensional folding of horizontal layers in a heterogeneous lithology as a response to constrictional strain (Fossen, et al., 2013), most clearly developed within with ENE-WSW striking Hwidiem Shear Zone and within the Kukuom-Juaboso Domain. Within the Ketesso Shear Zone, a quartz dioritic garnet amphibolite preserves an L>>S tectonite fabric (SB294; Fig. 3.4g) consistent with transtension in the lower crust (Dewey, 2002). It is juxtaposed with a weakly deformed, greenschist facies metabasalt (SB347) of the Sefwi-Wiawso Domain 13 km to the east-southeast. Differential exhumation juxtaposing highgrade tectono-metamorphic domains is attributed the normal movement along NNE-SSW-striking detachments faults and normal, left-lateral reactivation of major NE-striking shear zones, providing

evidence against homogeneous uplift proposed for Palaeoproterozoic orogenic domains involving weak, juvenile crust (Cagnard, *et al.*, 2006; Cagnard, *et al.*, 2007; Chardon, *et al.*, 2009; Cruden, *et al.*, 2006).

There are two possible interpretations of the driving force of transtension and constrictional strain. Within the hot orogen model, lateral constrictional flow develops during orogen-normal shortening in response to gravitational forces acting on thickened, hot lithosphere (Chardon, et al., 2009; Chardon, et al., 2011; Cruden, et al., 2006). Significant topographic relief and heterogeneous crustal thickening are prohibited by rheologically weak lithosphere and thermally homogeneous lower crust, with lateral shortening resulting in orogen-parallel flow, noted in the later stages of Neoarchaean convergence in the Gawler and Terre Adélie cratons (Duclaux, et al., 2007). A similar concept is applied to wide, mature collision orogens that have undergone extensive partial melting in the low crust, where accumulated gravitational stress results in extensional collapse (e.g. Brown, 2010; Royden, et al., 2008; Vanderhaeghe & Teyssier, 2001) and the formation of gneiss domes (Whitney, et al., 2004). It is likely that partial melting of middle and lower crustal rocks documented in the northeast of the study area contributed to the localised rheological weakening of the crust. However, the observed metamorphic textures and calculated metamorphic *P*-*T* paths do not display isothermal decompression, with steep retrograde paths instead indicating concurrent cooling and exhumation, therefore we do not consider gravitational collapse to be the primary driver of transtension-related exhumation. For SW Ghana, we hypothesise that constrictional deformation and transtension-driven late-orogenic exhumation indicates a change in boundary conditions to ~ENE-WSW divergence. This thermo-mechanical behaviour is analogous to the exhumation of HP and UHP rocks of the Western Gneiss Region of western Norway, driven by sinistral oblique plate divergence of Laurentian and Baltic plates (Krabbendam & Dewey, 1998).

3.7.3 Orogenic model, regional correlations and Transamazonian ties

Some of the major complexities when considering an orogenic model for the West African Craton arise from the regional or craton-scale correlation of metamorphic or deformation events between discrete greenstone belts, both volcano-plutonic and meta-sedimentary, and high-grade gneissic terranes. Polycyclic orogenic models for the West African Craton are inferred from exposures of high-grade terranes, thought to represent a Palaeoproterozoic (> 2150 Ma) basement deformed and metamorphosed prior the low-grade metamorphism and transcurrent tectonics of the Eburnean Orogeny (e.g. de Kock, *et al.*, 2012; Hein, 2010; Pouclet, *et al.*, 1996). In SW Ghana, D_1 deformation in both the Ashanti Belt (Perrouty, *et al.*, 2012) and the Sefwi Belt (this study) is associated with an initial phase of N-S to NNW-SSE shortening. Whilst we were unable to directly constrain the timing of D_1 in this study, Perrouty, *et al.* (2012) propose an Eoeburnean (D1) event between 2187 – 2158 Ma, observed in metavolcanic rocks and mica schists of the Sefwi Group exposed in the southern extent of the Ashanti Belt. This timing corresponds with extensive magmatism in the Sefwi Belt, regarded by other authors as a magmatic accretionary event (Allibone, *et al.*, 2002b; Feybesse, *et al.*, 2006; Milési, *et al.*, 1992). Perrouty, *et al.* (2012) bracket a subsequent extensional event associated with deposition of Kumasi Group volcano-sedimentary units between 2154 and 2125 Ma, prior to the deposition of Tarkwaian sediments between 2107–2097 Ma.

Correlation of the regional scale thrust faulting and crustal thickening documented in the Sefwi Belt, also associated with N-S shortening, is convoluted by the tectonic setting and timing of the deposition of Kumasi Group interpreted by Perrouty, *et al.* (2012). Rocks in the Chiraa and Kukuom-Juaboso domains indicate burial or supracrustal rocks to depths of >33 km, interpreted as a narrow collisional front between the Sefwi Belt and the Sunyani-Comoé Domain. The obliquity between D1 shortening and the NE-SW strike of the Sefwi Belt potentially suggests oblique collision along pre-existing crustal discontinuities associated with juvenile crust formation. Feybesse, *et al.* (2006) describe a similar period of thrust tectonism and crustal thickening between 2130 and 2105 Ma in SW Ghana, which they attribute to D_1 deformation of the Eburnean Orogeny. Such deformation may also provide a mechanism for the formation of shear-related, bedding-parallel foliations documented in both the Sefwi Belt and the Ashanti Belt (this study; Allibone, *et al.*, 2002b).

Poor correlation is found between subsequent deformation events between the Sefwi and the Ashanti belts. D_2 constrictional deformation described in this study is attributed to ENE-WSW transtension, normal left-lateral movement along major shear zones and a change in boundary conditions. Sinistral movements within the Sefwi shear systems is supported by field observations of Feybesse, *et al.* (2006) and in the Ivorian extent of the Sunyani-Comoé Domain (Vidal, *et al.*, 2009). *In situ* monazite ages constrain the initiation of D_2 exhumation of high-grade rocks at ca. 2073 Ma (Ch. 4; this study), post-dating granite and leucogranite emplacement in the Sunyani-Comoé domain between 2093 and 2081 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007), attributed to partial melting in the lower crust after to D_1 crustal thickening. We suggest that deposition of the Tarkwa-like immature conglomerates and sandstone along the NNE-striking Bibiani Shear zone is associated with graben formation and normal fault movement during D_2 . As such, we caution the correlation of these conglomerates with those exposed in the Ashanti Belt. Whilst D_2 potentially corresponds with the D_4 NNW-SSE shortening event responsible for the sinistral reactivation of the Ashanti fault (Allibone, *et al.*, 2002b; Perrouty, *et al.*, 2012), we are unable to reconcile the D_3 NW-SE shortening of Perrouty, *et al.* (2012) with deformation recorded in the study area. In addition, late dextral reactivation of regional NE-striking shear zones is not observed in the Ashanti Belt.

In NW Ghana, high-grade metamorphism develops during N-S shortening between 2140 and 2130 Ma (Block, et al., 2015; Block, et al., 2016). Exhumation of the lower crust is attributed to N-S extension during gravitational collapse generating anatectic domes. Block, et al. (2016) hypothesise a subsequent collision and corresponding E-W shortening between independent crustal domains represented by NW Ghana (D3; Block, et al., 2016) and SW Burkina Faso (D1; Baratoux, et al., 2011) between 2130 and 2110 Ma, forming a cohesive block. Anatexis and high-grade metamorphism in NW Ghana commence ~40 million years prior to the genesis of collision-related leucogranites along the NW margin of the Sefwi Belt after ca. 2092 Ma, indicating a diachronous metamorphic history for the high-grade terranes exposed in SW Ghana. Furthermore, the exhumation of middle and lower crust documented in both NW Ghana and the study area occur more than 40 million years apart. Such large discrepancies in the timing of deformation and metamorphism between discrete greenstone belts in SW Ghana relative to greenstone-granite-gneiss terranes of NW Ghana and SW Burkina Faso is therefore interpreted as the product of episodic collisional orogenesis between independent juvenile Palaeoproterozoic terranes. The proposed collisional D₁ event associated with NNW-SSE shortening in SW Ghana potentially corresponds with the post-2110 Ma late sinistral shearing along the N-S striking Jirapa shear zone of NW Ghana (Block, et al., 2016). Sinistral shear zones in NW Ghana are subsequently overprinted by dextral NE-striking strike-slip shear zones during E-W shortening (D5; Block, et al., 2016), which we interpret as synchronous with our D₃ deformation. Such deformation may be the distal expression of the collision of the Archaean Kénéma-Man Domain and the Baoulé-Mossi domain between 2050 and 2030 Ma (Kouamelan, et al., 1997; Pitra, et al., 2010).

On a more global scale, geological, structural and geochronological studies employed in paleogeographic and geochronological reconstructions have supplied strong evidence that West Africa and the São Luís Craton were contiguous during the Palaeoproterozoic (e.g. Hurley, *et al.*, 1967; Klein & Moura, 2008, and references therein; Torquato & Cordani, 1981). Magmatism of the São Luís craton and the basement of the Gurupi belt is defined by three periods of magmatic activity (2240 \pm 5 Ma; 2160 \pm 10 Ma; 2080 \pm 20 Ma), sharing lithological and structural characteristics with SW Ghana (Klein & Moura, 2008). Peraluminous collision-type granitoids (2086 – 2090 Ma), produced by partial melting of crustal material during collision (Klein, et al., 2005; Klein & Moura, 2008), are of comparable age to the collisional orogenesis and leucogranites emplacement in the Sunyani-Comoé Domain (Agyei Duodu, et al., 2009; Hirdes, et al., 1992; Hirdes, et al., 2007). A stronger correlation may be made between the magmatic and tectono-thermal evolution of the Sefwi Greenstone Belt and the Ile de Cayenne complex, French Guiana. Both the Sefwi Belt and the Cayenne complex are characterised by early tholeiitic magmatism followed by emplacement of TTG magmas between ca. 2174 and 2144 Ma, with some bearing calc-alkaline affinities (Delor, et al., 2003; Feybesse, et al., 2006; Hirdes, *et al.*, 2007; Vanderhaeghe, *et al.*, 1998). Both terranes yield positive $\epsilon Hf_{(t)}$ and $\epsilon Nd_{(t)}$ values that attest to the juvenile character of the crust (Delor, et al., 2003; Petersson, et al., 2017; Taylor, et al., 1992), which formed distal to Archaean cratonic material with a subsequent period tectonic accretion (Gruau, et al., 1985; Kroonenberg, et al.; Vanderhaeghe, et al., 1998). Crustal thickening, anatexis and emplacement of potassic granitoids and peraluminous leucogranites along major shear zones between ca. 2093 and 2083 Ma in French Guiana is attributed to oblique collision and late orogenic lateral extrusion (Vanderhaeghe, et al., 1998), bearing a striking similarity to the sequence of events recorded in southwest Ghana. The final correlation between the two terranes comes from the timing of crustal extension and formation of granulite facies metamorphism is in the Guiana Shield at 2070 to 2060 Ma (Delor, et al., 2001), which we interpret to be a lateral continuation of the D₂ deformation and metamorphism recorded on the northwest margin of the Sefwi Belt. Furthermore, if we assign sinistral movement, as documented in this study, to the 400km of displacement along the Sefwi shear system proposed by Jessell, et al. (2012), it would place modern day SW Ghana, and likely the São Luís Craton, in NE Brazil in the French Guiana Shield.

3.7.4 Implications for Paleoproterozoic geodynamics

Structural and metamorphic evidence yielded by rocks in the West African Craton provide an opportunity to explore orogenic processes during the Palaeoproterozoic, prior to the appearance of blueschists and ultra-high-pressure metamorphism (UHPM) and other tectonic indicators attributed to a modern plate tectonic regime in the Neoproterozoic (Stern, 2005; Stern, 2007). The Neoarchaean to Palaeoproterozoic period, however, coincides with a number of significant changes in the Earth, including changes in the tectonic style, secular evolution recorded in the metamorphic record and

evolving chemistry of magmatic rocks (e.g. Brown, 2006; Brown, 2007; Dhuime, *et al.*, 2012; Keller & Schoene, 2012; Laurent, *et al.*, 2014). Indeed, it is widely accepted that the ambient upper-mantle temperatures were hotter in the past (e.g. Herzberg, *et al.*, 2010; Labrosse & Jaupart, 2007), likely affecting the tectonic style and the rheology of the lithosphere (Burov & Yamato, 2008; Burov, 2011; Sizova, *et al.*, 2010; van Hunen & van den Berg, 2008). The orogenic model we propose for SW Ghana reveals structural and metamorphic evidence of significant crustal thickening and clockwise *P-T-t* paths reminiscent of modern collisional orogens, forming a growing body of research suggesting collision between discrete Palaeoproterozoic terranes in the West African Craton (Block, *et al.*, 2016; Parra-Avila, 2015; Parra-Avila, *et al.*, 2016).

We iterate that the *hot-orogen* model does not sufficiently explain the strain patterns and metamorphism documented in southwest Ghana which contrast the homogeneous lithospheric deformation, monotonous metamorphism and limited exhumation capacity of hot orogens proposed for Neoarchaean and Palaeoproterozoic orogens (Cagnard, *et al.*, 2011; Cagnard, *et al.*, 2006; Cagnard, *et al.*, 2007; Chardon, *et al.*, 2009; Rey & Houseman, 2006; Rey, *et al.*, 2003), proposed also for areas of the West African Craton (Vidal, *et al.*, 2009). Instead, the tectonic style and orogenic model proposed for SW Ghana bears a number of similarities to collisional orogenesis documented in the Archaean Barberton terrain (Diener, *et al.*, 2013; Moyen, *et al.*, 2006) and the Archaean Abitibi Subprovince (Daigneault, *et al.*, 2002; Mueller, *et al.*, 1996), including initial HP-MT metamorphism interpreted as a collisional event followed by late orogenic extension.

The period of global orogenesis between ca. 2100 and 1800 Ma is characterised by extremely diverse metamorphic conditions and tectonic style including the pene-contemporaneous 1810 Ma hot orogen of the southern Finnish Svecofennides (Cagnard, *et al.*, 2007) and the Himalaya-like Trans-Hudson Orogen, associated with the formation and exhumation of eclogite facies rocks (Weller & St-Onge, 2017). As such, we believe the tectono-metamorphic history of SW Ghana supports the notion of non-unique orogenic processes for the Neoarchaean-Palaeoproterozoic period, potentially reflecting the thermal divergence in the mantle between 2.500 and 2000 Ma, as proposed by Condie, *et al.* (2016) and the progressive, and possibly localised, development of plate tectonics and subduction settings (Condie & Kröner, 2008).

3.8 Conclusions

Ghana during the Palaeoproterozoic Eburnean Orogeny has been presented, based on the integration of regional field mapping, interpretation of regional airborne geophysical datasets and incorporation of geochronological constraints. The re-interpretation of the litho-tectonic map, combined with the regional structural framework and complementary metamorphic data suggests that the domains within the Sefwi Belt record significant crustal thickening during NNW-SSE shortening, relative to adjacent low-grade domains. The tectonic contact between the Sefwi Belt and the Sunyani-Comoé Domain is interpreted as a narrow, collisional front, representing a suture zone between SW Ghana and central Ghana/Ivory Coast that formed during oblique convergence. Differential exhumation of high-grade domains occurred after thermal relaxation of the orogenic crust during D₂ transtension, contrasting the surface geology predicted for Precambrian hot orogens. Furthermore, based on previous studies of elongate granites coupled with macro-geometries of domains, we interpret major normal, left lateral movement within the Sefwi shear system. Based on both structural and metamorphic evidence transtension is interpreted as the product of orogenic extension reflecting changes in boundary conditions and oblique plate divergence rather than gravitational collapse. Localised, late dextral reactivation of NE-striking shear zones is generated by far-field stresses. We highlight the diachronous nature of high-grade metamorphism and discrepancies in deformation histories across the craton during the Eburnean Orogeny, which we interpret as the product of episodic collisional orogenesis between discrete juvenile terranes during the assembly and stabilisation of the West African Craton.

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Appendix A. Supplementary data I – IV

See pp. 276 - 279 or refer to electronic data file "Ch3_AppendixA_SI"

Fig. A1SI. Composite image of greyscale reduced to the pole (RTP) of the residual magnetic intensity and associated individual survey details.

Fig. A2SI. Bouguer gravity anomaly grid of survey area

Fig. A3SI. Composite image of negative reduced to the equator (RTE) of the residual magnetic intensity (RMI) with greyscale first vertical derivative (1VD) for all surveys utilised in the study

Fig. A4SI. Greyscale digital elevation model of study area with ternary image of gamma-ray signal from three channels, K, eTh, and eU, highlighting the partial coverage of radiometric data of the study area

Appendix B. Magnetic susceptibility histograms

See pp. 280 or refer to electronic data file "Ch3_AppendixB"

Fig. B1. Measured magnetic susceptibility values for individual units presented in histograms to display range and frequency.

Chapter IV

The metamorphic evolution of SW Ghana: Insights into the Palaeoproterozoic Eburnean Orogeny, West African Craton



4.0 Introduction

This chapter is complementary to the tectono-metamorphic framework and map presented in Chapter 3, providing detailed petrographic analysis, quantitative estimates of prevailing metamorphic conditions and geochronological constraints on deformation and metamorphism during the Eburnean Orogeny. These new geochronological constraints are compared to existing geochronological data in order to understand the timing of metamorphism in high- and low-grade terranes, as well as the timing of high-grade metamorphism in the study area relative to other high-grade terranes within the craton. The metamorphic assemblages and monazite ages documented in this chapter elucidate a more complex metamorphic history for the SW Ghana and the Eburnean Orogeny, thus aiding our understanding of the amalgamation of the West African Craton. We consider the significance of the metamorphism documented in SW Ghana relative to the global metamorphic record, which appears to be unique the Palaeoproterozoic plate tectonic regime.

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Chapter 4: Petrological and geochronological evidence of collisional orogenesis and lower crust exhumation during the Palaeoproterozoic Eburnean Orogeny, SW Ghana, West African Craton

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Abstract

New petrological and geochronological data are presented for migmatitic paragneisses and high-grade metavolcanic rocks exposed in southwest Ghana in the south-eastern portion of the Palaeoproterozoic (2300–2070 Ma) West African Craton. The study focuses on tectono-metamorphic domains within and adjacent to the NE-striking Sefwi Greenstone Belt, formed during the relatively understudied Eburnean Orogeny (2150–2070 Ma). These domains are bounded or transected by steeply dipping NE- to ENE-striking reactivated regional ductile shear zones.

North of the Sefwi Belt, a newly interpreted high-grade terrane comprises heterogeneous, migmatitic paragneisses with voluminous biotite-muscovite granite intrusions and is interpreted to be part of the Sefwi Belt due to its metamorphic and structural similarities. Paragneiss and garnet amphibolite samples sporadically reveal initial high-pressure amphibolite facies metamorphic conditions (10.0-11.5 kbar, 580-650 °C), corresponding to a cool apparent geothermal gradient of 15-17 °Ckm⁻¹. Migmatites record peak temperature conditions at the amphibolite-granulite facies transition (7.5–9.5 kbar, 650–700 °C), establishing a new apparent geothermal gradient of 20–25 °C km⁻¹. Variable prograde paths in the migmatites indicate contrasting prograde paths showing heating during decompression, or, conversely, concomitant heating and burial. Subsequent coeval exhumation is suggested by parallel clockwise P-T-t paths, simultaneous decreases in pressure and temperature and a regional amphibolite facies metamorphic overprint. In-situ monazite U-Pb SHRIMP ages of ca. 2073 Ma constrain the timing of the initial stages of cooling and exhumation. In contrast to the migmatites, a garnet amphibolite sample records cooling during exhumation immediately following initial HP-MT conditions. We interpret the clockwise P-T-t paths of two of the three samples as evidence of convergence-related crustal thickening and collision along the northwest margin of the Sefwi greenstone belt. Exhumation of the lower crust along NE-striking shear zones, and west-dipping normal detachment post-dates leucogranite emplacement in lower and upper crustal slices at ca. 2092 – 2081 Ma.

High-pressure metamorphic assemblages and low apparent geothermal gradients distinguish SW Ghana from similar Precambrian granite-greenstone provinces elsewhere. Furthermore, diverse thermal histories and the tectonic juxtaposition of crustal slices from different depths do not conform with the homogeneous metamorphism and strain patterns attributed to the "hot orogen" model proposed for Precambrian orogens. Furthermore, such young metamorphic ages have not been previously documented in the south-eastern portion of the West African Craton. These findings instead suggest episodic collisional orogenesis occurred during craton assembly resulting in a diachronous metamorphic history. The metamorphic record of the West African Craton supports the notion of a Palaeoproterozoic *transitional* tectonic regime, prior to the appearance of the hallmark ultra-high-pressure (UHP) metamorphism attributed to modern subduction-related plate tectonics.

Key words: collision; exhumation; transtension; Palaeoproterozoic orogenesis; West African

Craton.

4.1 Introduction

Precambrian geodynamics and lithosphere tectonics continue to provide a source of controversy in scientific literature (e.g. Cawood, et al., 2006; Condie & Benn, 2006; Reddy & Evans, 2009). Of particular interest are the tectonic settings and processes associated with Precambrian accretionary and collisional orogenesis, culminating in craton assembly and stabilisation (Condie, 2007; Windley, 1992). Precambrian accretionary orogens include all Archaean cratons and several Palaeoproterozoic provinces (Windley, 1992) and are characterised by high crustal production rates, voluminous juvenile magmatism, elevated geothermal gradients and low lithospheric strength relative to the Phanerozoic (Condie, 2007; Gerya, 2014; Sizova, et al., 2010). The comparison of metamorphic conditions and tectonic settings preserved in Precambrian orogens relative to the Phanerozoic, therefore, remains uncertain. Archaean orogens commonly comprise elongate greenstone belts and vast granite-gneiss domains, featuring low-pressure granulite facies metamorphism and homogeneous strain patterns (e.g. Bouhallier, et al., 1995; Choukroune, et al., 1995). Whilst some authors interpret such features within the modern plate tectonics paradigm (e.g. Cawood, et al., 2006; de Wit, et al., 1992), alternative tectonic processes have also been proposed by a number of authors (e.g. Bouhallier, et al., 1995; Choukroune, et al., 1995; Hamilton, 1998; Rey & Houseman, 2006), emphasising the potential significance of diapirism, vertical movement and gravitational forces in granite-greenstone regions during lithospheric deformation. Additionally, the "hot orogen" model proposed for some Precambrian provinces accommodates lateral shortening in weak, hot crust through distributed deformation, with minimal topographic expression due to lateral or three-dimensional flow (Cagnard, et al., 2006; Cruden, et al., 2006; Rey & Houseman, 2006). Such orogens are commonly characterized by broad regions with isobaric metamorphism and lateral temperature gradients (Bleeker, 1990; Caby, et al., 2000; Cagnard, et al., 2007; Chardon, et al., 2009), resulting from homogeneous uplift of the entire domain (e.g. Gapais, et al., 2009). In contrast, Phanerozoic accretionary orogens involve subduction of cold, rigid lithosphere, able to support significant crustal thickening, thus generating and exhuming high-pressure (HP) to ultra-high-pressure (UHP) metamorphic terranes, inferred as the hallmark of modern-style plate tectonics (Stern, 2005; 2007).

A number of studies, however, have demonstrated the sporadic appearance in the Neoarchaean and Palaeoproterozoic of high-pressure and/or eclogite facies metamorphic conditions and diverse geothermal gradients within metamorphic domains of Precambrian orogens (e.g. Block, *et al.*, 2013; Collins, *et al.*, 2004; Ganne, *et al.*, 2012; Moyen, *et al.*, 2006; Nicoli, *et al.*, 2015; Volodichev, *et*

al., 2004). In a review of the secular evolution of metamorphism, Brown (2007; 2009) notes the appearance of complementary eclogite-high-pressure granulite (E-HPG) and ultra-high temperature (UHT) metamorphism, which he infers to be the ancient expression of subduction-to-collision orogenesis. Brown (2006) proposes a transitional "Proterozoic plate tectonic" regime based on the diversification of recorded metamorphic conditions after the Neoarchaean. Rare high P/T and low geothermal gradients (<20 °km⁻¹) and evidence of structurally controlled exhumation are documented in the MesoArchaean Barberton greenstone-granitoid terrane (Moyen, et al., 2006), the Neoarchaean to Palaeoproterozoic Jiaodong Terrain in the North China Craton (Liu, et al., 2015), the Salem block of southern India (Anderson, et al., 2012), the Usagaran Orogen, Tanzania (Collins, et al., 2004), the Fada N'Gourma area of Burkina Faso (Ganne, et al., 2012) and high-grade paragneisses in NW Ghana (Block, et al., 2015), where the latter two examples are from the West African Craton (WAC). However, debate continues to surround the tectonic setting producing HP-UHP metamorphic rocks and the mechanism(s) of exhumation in the Neoarchaean and Palaeoproterozoic (Block, et al., 2015; Block, et al., 2016b; Brown, 2007; Ganne, et al., 2014; van Hunen & van den Berg, 2008). In this paper, we investigate variations in the thermal history and exhumation mechanisms active during the Palaeoproterozoic Eburnean Orogeny of the WAC, focusing on newly documented high-grade domains of the Sefwi greenstone belt in southwest Ghana.

The Palaeoproterozoic Baoulé-Mossi domain (2300–2070 Ma) of the southern WAC is a vast (3 x 10⁶ km²) juvenile crustal domain, comprising Archaean-like tonalite-trondhjemite-granodiorite granite-gneiss domains and linear-to-arcuate metavolcanic and metasedimentary greenstone belts, bound by crustal-scale strike-slip shear zones (Bessoles, 1977). Widespread crustal reworking and the emplacement of more evolved, potassic granites is associated with terrane accretion and the collision of the juvenile Palaeoproterozoic domain with the Archaean Kénéma-Man nucleus during the 2150 – 2070 Ma Eburnean Orogeny (Bonhomme, 1962; Feybesse, *et al.*, 2006; Hirdes, *et al.*, 1996; Kouamelan, *et al.*, 1997; Ledru, *et al.*, 1994; Milési, *et al.*, 1991; Pitra, *et al.*, 2010). The Eburnean Orogeny represents one of the oldest Palaeoproterozoic orogens on Earth during a period of global orogenesis between 2100 Ma and 1800 Ma, and is likely to herald the onset of the amalgamation of the supercontinent, Nuna (or Columbia) (Rogers & Santosh, 2002; Zhao, *et al.*, 2002; Zhao, *et al.*, 2004).

There remains, however, disagreement regarding the tectonic processes responsible for the WAC assembly, which can be partially attributed to the limited number of studies with integrated metamorphic and geochronological data (e.g. Block, *et al.*, 2015; Feybesse, *et al.*, 2006; Kouamelan, *et al.*, 1997; Pitra,

et al., 2010). In this study we combine field observations, petrological and geochronological data to investigate spatial and temporal variation in the formation and exhumation of high-grade tectonometamorphic domains juxtaposed with low-grade supracrustal rocks in the south-eastern portion of the WAC. We address ambiguities in the relationship between high-grade gneissic domains, greenstone belts *sensu stricto* and metasedimentary provinces. The results are compared with those of Block *et al.* (2015) and we discuss the potential tectonic models proposed for the WAC. The findings have significant implications for understanding the enigmatic crustal architecture, including deciphering whether high-grade domains represent an older crustal basement or lower crustal equivalents of low-grade metavolcanic and metasedimentary supracrustal rocks.

4.2 Geological Setting

4.2.1 West African Craton geology

The West African Craton (WAC) consists of the northern Reguibat Rise, the Kédougou-Kéniéba and Kayes inliers and the southern Leo-Man Rise (Bessoles, 1977; Rocci, *et al.*, 1991), overlain by the Neoproterozoic to Devonian Taoudeni basin in the central part of the craton. The Leo-Man Rise (Fig. 4.1) comprises the western Archaean Kénéma-Man domain, flanked to the north and east by the Palaeoproterozoic Baoulé-Mossi domain, which are tectonically juxtaposed along the Sassandra Fault (Bessoles, 1977). Palaeoproterozoic rocks are characterised by elongate volcanic, volcano-sedimentary and sedimentary belts of the Birimian Supergroup and voluminous granite-gneiss terranes (Fig. 4.1). The Birimian Supergroup is subdivided into basaltic to rhyolitic volcanic and volcano-sedimentary rocks of the Sefwi Group (Perrouty, *et al.*, 2012) and the overlying volcaniclastic, volcano-sedimentary and rare chemical (Mn-rich) sedimentary rocks of the Kumasi Group (Adadey, *et al.*, 2009; Perrouty, *et al.*, 2012). Volcanic units display tholeiitic to calc-alkaline affinities (e.g. Abouchami, *et al.*, 1990; Baratoux, *et al.*, 1998; Hirdes & Davis, 1998; Hirdes, *et al.*, 1996; Loh, *et al.*, 1999; Pouclet, *et al.*, 2006).

Deposition of the Birimian volcano-sedimentary units is constrained by detrital zircon, indicating a maximum deposition age of ca. 2150 Ma (Davis, *et al.*, 1994; Oberthür, *et al.*, 1998), continuing until late-depositional emplacement of a granodiorite pluton in the Kumasi Basin at 2136 \pm 19 (Adadey, *et al.*, 2009). Volcano-plutonic greenstone belts are commonly unconformably overlain by shallow water, fluvio-deltaic sedimentary rocks of the Tarkwaian Group (Kitson, 1918; Whitelaw & Junner, 1929) deposited in small fault-bounded basins, with detrital zircons constraining a maximum deposition age of c. 2130, or possibly as young as c. 2107 Ma (Davis, *et al.*, 1994; Hirdes & Nunoo, 1994; Pigois, *et al.*, 2003).



Fig. 4.1. Geological map of the Leo-Man rise (modified after BRGM SIGAfrique, Milési, et al., 2004). Mafic to intermediate volcanic greenstone belts (dark green); volcano-sedimentary, volcaniclastic and intermediate volcanic provinces (light green); Granitoid composition ranges from tonalite-trondhjemite-granodiorite and granite with minor mafic intrusions. Late fluvio-deltaic sediments (yellow) unconformably overlying greenstone belts in narrow, fault bounded basins.

Granite-gneiss terranes of the West Africa Craton formed during several protracted pulses of intense Na₂O- and CaO-rich trondhjemite-tonalite-granodiorite (TTG) plutonic activity between ca. 2200 and 2130 Ma (Doumbia, *et al.*, 1998; Egal, *et al.*, 2002; Feybesse, *et al.*, 2006; Gasquet, *et al.*, 2003; Hirdes, *et al.*, 1996; Oberthür, *et al.*, 1998). Subsequent magmatism evolved towards potassic, peraluminous granites by ca. 2070 Ma (e.g. Block, *et al.*, 2016a; Egal, *et al.*, 2002; Gueye, *et al.*, 2008; Hirdes, *et al.*, 1992; Lompo, 2009).

Initial juvenile crustal formation is speculated to have evolved from an ocean plateau in association with mantle plume activity, with subsequent calc-alkaline plutonism and contemporaneous volcanism (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992; Leube, *et al.*, 1990) or alternatively from arc volcanism followed by micro-plate accretion (Baratoux, *et al.*, 2011; Dampare, *et al.*, 2008; Petersson,

et al., 2016; Pouclet, *et al.*, 1996; Sylvester & Attoh, 1992). This juvenile material was deformed during the c. 2150-1980 Ma Eburnean Orogeny (Bonhomme, 1962). Early crustal thickening was attributed by some authors to magmatic accretion similar to a modern plate tectonic setting, which was subsequently overprinted by thrust faults (e.g. Feybesse, *et al.*, 2006; Feybesse & Milési, 1994; Milési, *et al.*, 1992). Conversely, "Archaic-type" vertical tectonics that emphasised pluton diapirism and the relative downward movement of greenstone belts has also been proposed (Delor, *et al.*, 1995; Pouclet, *et al.*, 1996; Vidal, *et al.*, 1996; Vidal, *et al.*, 2009). Across the craton, regional metamorphism associated with the Eburnean Orogeny varies from lower-greenschist facies to lower-amphibolite facies (e.g. Béziat, *et al.*, 2000; Feybesse, *et al.*, 2006; Ganne, *et al.*, 2012; Kříbek, *et al.*, 2008). Upper-amphibolite facies conditions are generally restricted to the contact aureoles of plutons (Debat, *et al.*, 2003; Lobjoit, 1964; Pons, *et al.*, 1995). The absence of HP-LT assemblages and significant lateral metamorphic breaks has been attributed by some authors to elevated geothermal gradients (>25 °Ckm⁻¹) and homogenous uplift (Pons, *et al.*, 1995; Vidal, *et al.*, 2009).

Recent studies, however, have revealed heterogeneity in thermal gradients and metamorphic conditions (Block, et al., 2015; Ganne, et al., 2012). In southern Ghana, regional MP-MT amphibolite facies metamorphism (5-6 kbar, 500-600 °C) is recorded within the Sefwi and Ashanti belts (Galipp, et al., 2003; John, et al., 1999), and slightly higher anatectic conditions occur in the Winneba-Awutu migmatites on the Ghanaian southern coast (>5 kbar, >600 °C) (Loh, et al., 1999; Opare-Addo, et al., 1993). High-pressure granulites are restricted to the tectonic contact between the Archaean Kénéma-Man shield and the Palaeoproterozoic Baoulé-Mossi domain (Fig. 4.1), where conditions reached ~13 kbar and 850 °C (Pitra, et al., 2010). Ganne, et al. (2012) documented an early, cold thermal regime (<10–15 °Ckm⁻¹) and high-P greenschist to blueschist facies conditions in chlorite-phengite assemblages in greenstone belts in eastern Burkina Faso. Similar relic metamorphic conditions (8-10 kbar, 400-500 °C and 11-14 kbar, 520-560 °C) are preserved in volcano-sedimentary schists and paragneisses of northwest Ghana (Block, et al., 2015). Where observed, sharp metamorphic breaks between high-and low-grade domain are interpreted as the product of tectonic exhumation and juxtaposition of different crustal levels (Block, et al., 2015; Block, et al., 2016b; Hirdes, et al., 2007; Loh, et al., 1999; Opare-Addo, et al., 1993), which contradicts lateral, granitoid-related metamorphic gradients used to interpret high- and low-grade domains within the "hot" orogen models (Chardon, et al., 2009; Vidal, et al., 2009).

There are a number of models proposed for the tectono-metamorphic evolution of the West

African Craton based on the interpretation of the relationship between high-grade metamorphic domains and low-grade volcano-sedimentary sequences. Small exposures of poly-deformed high-grade rocks in Burkina Faso and Ivory Coast (Boher, *et al.*, 1992; de Kock, *et al.*, 2011; 2012), or poly-deformed metavolcanics in SW Ghana (Perrouty, *et al.*, 2012), have been interpreted as the basement upon which Birimian metasedimentary sequences were deposited. This sporadically recognised event is referred to as Eburnean I (Allibone, *et al.*, 2002) or Eoeburnean (de Kock, *et al.*, 2011; Perrouty, *et al.*, 2012) in Ghana and western Burkina Faso (Baratoux, *et al.*, 2011; Metelka, *et al.*, 2011), and the Tangaean event in north-eastern Burkina Faso (Hein, 2010; Tshibubudze, *et al.*, 2009). It precedes the more widely recognised phase of the Eburnean (or Eburnean II) Orogeny (~2130 – 1980 Ma), associated greenschist facies metamorphism and transcurrent tectonism (Allibone, *et al.*, 2002; Eisenlohr & Hirdes, 1992; Feybesse, *et al.*, 2006; Jessell, *et al.*, 2012; Pouclet, *et al.*, 2006; Pouclet, *et al.*, 1996; Vidal, *et al.*, 2009). Alternatively, a monocyclic evolution is proposed for the Eburnean Orogeny, during which high-grade gneisses and migmatites and low-grade rocks undergo coeval metamorphism at different crustal levels during a single progressive event and are subsequently tectonically juxtaposed (Block, *et al.*, 2015; Block, *et al.*, 2016); Eisenlohr & Hirdes, 1992; Opare-Addo, *et al.*, 2016);

Distinguishing between monocyclic and polycyclic models for the Eburnean orogeny is hampered by the limited amount of precise geochronological data constraining the time of high-grade metamorphic events, most notably lacking in southern Ghana. Amphibolite facies metamorphism in central Ivory Coast is dated at 2153 ± 93 Ma (Grt-WR Sm-Nd; Boher, et al., 1992) and 2144 ± 3 Ma in the Ivorian extent of the Sefwi belt (Pb-Pb Tnt; Hirdes, et al., 2007). In-situ ICP-MS and SHRIMP U-Pb monazite crystallisation ages, hosted by leucosomes and overprinting assemblages constrained anatexis at 2138 ± 7 Ma and overprinting amphibolite facies metamorphism at 2130 ± 7 Ma in migmatitic para- and orthogneissic units in NW Ghana (Block, et al., 2015). These ages are distinctly older than geochronological data revealed by syntectonic intrusions and metamorphic units from across the craton, which constrain Eburnean deformation and low-grade metamorphism between 2114 and 2080 Ma (de Kock, et al., 2011; Feybesse, et al., 2006; Hirdes, et al., 1996; Kouamelan, et al., 1997). The final tectono-metamorphic event records the collision of the Archaean and Palaeoproterozoic domains between ~2080 and 2030 Ma based on U-Pb zircon, monazite, titanite ages and Sm-Nd garnet ages (Kouamelan, et al., 1997; Pitra, et al., 2010). Whilst disparity in metamorphic ages recorded across the craton might suggest multiple metamorphic peaks between ca. 2150 and 2030 Ma, further integrated tectono-metamorphic and geochronological data are required to unravel the tectonic evolution and assembly of the WAC.

4.2.2 Geology of southwest Ghana

The Palaeoproterozoic rocks of southern Ghana and southeast Ivory Coast are characterised by NE-striking, parallel greenstone belts several hundred kilometres long and 20–60 km wide, with large, intervening metasedimentary domains. This study focuses on the fault-bounded, lithotectonic domains of the Sefwi greenstone belt and the adjacent Sunyani-Comoé and Kumasi-Afema metasedimentary domains (Fig. 4.2). Outcrop is exceedingly sparse and the field area is extensively covered by thick regolith, cocoa plantations and forest reserves.

The field area is dominated by the 280 km-long, 40-80 km-wide Sefwi Greenstone Belt, which is bounded to the north by the Kenyase-Yamfo shear zone and to the south by the Afema shear zone (Hirdes, et al., 2007) (Fig. 4.2). It is transected by the Ketesso High Shear Zone, which trends parallel to the Sefwi Belt structural grain (Hirdes, et al., 2007; Jessell, et al., 2012). In this study, the belt is subdivided into four litho-tectonic domains: the Kukuom-Juaboso domain (KJD), the Bechem domain (BD), the Hwidiem domain (HD) and the southern Sefwi-Wiawso domain (SWD) (Fig. 4.2). The central, rhomboidal Kukuom-Juaboso domain, is an 80 km long, elongate in a NEtrending domain that comprises amphibolite facies mafic to intermediate volcanic rocks, immature volcaniclastic greywackes, minor metasedimentary units and extensive granitoids intrusions (Galipp, et al., 2003; Hirdes, et al., 1993). The Bechem granitoid domain is characterised by layered intrusions of gabbro, norite and pyroxenite, strongly deformed amphibolites, large quartz-diorite and granodiorite plutons, and small, elliptical post-tectonic granites (Agyei Duodu, et al., 2009). The northern margin of the adjacent Bechem and Kukuom-Juaboso domains is marked by a ENE-striking shear zone, which contains slivers of migmatitic mafic orthogneisses (this study). Geochronological constraints for magmatism for these domains are limited to a rhyolite yielding a U-Pb zircon age of 2189 ± 1 Ma (Hirdes & Davis, 1998) and a hornblende-biotite granodiorite U-Pb zircon age of 2135 Ma (Amponsah, 2012). The NE-striking Ketesso shear zone transects the central region of the Sefwi belt, south of which is the low-grade Sefwi-Wiawso volcano-plutonic domain of the belt, comprising greenschist facies mafic to intermediate lavas and abundant granitoid intrusions that display pervasive chlorite-carbonate or quartz-sericite alteration with minor accessory sulphides. U-Pb zircon ages for granitoids in the Sefwi-Wiawso domain yield crystallisation ages between 2180 and 2140 Ma (Hirdes, et al., 1992; Hirdes, et al., 2007).



Fig. 4.2. Regional litho-structural map of the study area (Chapter 4: this study). The Sunyani-Comoé and Kumasi-Afema domains are characterised by isoclinally folded, sub-greenschist to greenschist facies metasedimentary and volcano-sedimentary rocks with rare intercalated intermediate volcanic layers. The Chiraa domain (CD) in the northeast comprises poly-deformed paragneisses and voluminous granite intrusions. Lithologies exposed in the Kukuom-Juaboso domain (KJD) are diverse, consisting of metamorphosed mafic to intermediate volcanic, volcaniclastic and minor sedimentary units and multiple generations of felsic and mafic granitoid intrusions. BD: Bechem domain; SWD: Sefwi-Wiawso domain; HD: Hwidiem domain; HSZ: Hwidiem shear zone. Fault kinematics denoted only where unambiguous, or, in the case of the Ashanti belt, derived from the literature (e.g. Perrouty, et al., 2012). U-Pb zircons ages are sourced from: (1) Petersson, et al. (2016); (2) Hirdes, et al. (1992); (3) Delor, et al. (1992).

Towards the northeast of the study area, the Chiraa high-grade domain (formerly referred to as the Sunyani batholith) is a 25 x 35 km triangular shaped domain (Fig. 4.2), comprising strongly deformed, migmatitic paragneiss with abundant granite intrusions and pegmatite dykes, amphibolite enclaves and rare migmatitic mafic orthogneisses. Granitic intrusions often contain tectonic slivers of paragneiss and biotite-rich schlerien. Large exposures of paragneiss units comprise alternating pelitic and psammopelitic layers. Rare exposures of felsic orthogneiss display a gneissosity parallel with that of the paragneisses. This is in tectonic contact with greenschist facies volcano-sedimentary rocks of the Sunyani-Comoé basin along a NNE-striking fault. The eastern margin of the Chiraa high-

grade domain is unconformably overlain by undeformed Neoproterozoic Voltaian sedimentary rocks (Kalsbeek, *et al.*, 2008).

The Sunyani-Comoé Domain comprises low-grade chlorite-white mica schists, greywackes, argillites, banded iron stones, graphitic shales, volcaniclastic rocks and rare intercalated intermediate volcanic rocks of the Kumasi Group (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 2007; Hirdes, *et al.*, 1993). Similar lithologies are found in the Kumasi-Afema Domain to the south of the study area, south of which lies the Ashanti Greenstone Belt (Agyei Duodu, *et al.*, 2009; Perrouty, *et al.*, 2012). U-Pb zircon crystallization ages (ca. 2093 – 2081 Ma) of leucogranites intruding the Sunyani-Comoé Domain parallel to the north-western margin of the Sefwi Belt (Fig. 4.2) (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 2007; Hirdes, *et al.*, 1993; Petersson, *et al.*, 2016).

4.3 Methods

4.3.1 Petrographic analysis and mineral chemistry

Mineral assemblages, textural context and microstructures were examined in polished thin sections under transmitted, plane polarised (PPL) and cross polarised light (XPL). Backscattered electron (BSE) imaging was performed at Monash Centre for Electron Microscopy at Monash University, Australia, using a JEOL JSM-7001F Field Emission Gun (FEG) Scanning Electron Microscope (SEM). Preliminary mineral chemistry examined by energy-dispersive x-ray spectroscopy (EDS) using Oxford Instruments Aztec X-ray analysis system with 80mm² Silicon Drift Detector (SDD). Conditions used for BSE and EDS mineral chemical composition characterization are: accelerating voltage of 15 keV, working distance (WD) of 10mm, and probe current of 10A, objective aperture of 3. To image internal structures of monazites, accelerating voltage was increased to 20 keV with a probe current of 14. Further BSE imaging and chemical characterization was performed at Géoscience Environnement Toulouse, Université Paul Sabatier (Toulouse III), France, using a JEOL JSM 6360LV coupled to a PGT Sahara SDD EDS detector. Imaging conditions used accelerating voltage of 20 keV, WD of 10mm and spot size of 50µm. Mineral compositions were measured using a Cameca-SX-5 at Centre de Microcaractérisation Raimond CASTAING in Toulouse, France.

4.3.2 *P-T* calculations

To derive P-T conditions and metamorphic paths, pseudosections were calculated for the whole rock compositions, calculating minimum Gibbs free energy in the P-T space represented by a multiphase

equilibrium diagram. Pseudosections were constructed using the computer software package Perple_X (version 6.6.8) (Connolly, 1990; Connolly, 2005) (http://www.perplex.ethz.ch/). Calculations were conducted in the Na₂O-K₂O-CaO-MgO-MnO-FeO-Al₂O₃-TiO₂-SiO₂-H₂O chemical system for metapelitic samples and in the Na₂O-CaO-MgO-MnO-FeO-Al₂O₃-TiO₂-SiO₂-H₂O for mafic samples. The bulk composition of each sample was obtained with X-ray fluorescence (XRF) at the Australian Research funded Centre of Excellence in Ore Deposits (CODES) at the University of Tasmania using a PANalytical Axios Advanced X-Ray Spectrometer (Robinson, 2003; Watson, 1996). CaO was corrected for apatite content using measured P_2O_5 , and ferric iron was transformed into ferrous iron using the following equation:

M Fetotal = M FeO (wt%) = $0.9 \times M$ Fe2O3 (wt %).

The internally consistent thermodynamic dataset of Holland and Powell (1998) (updated in 2002) was used for minerals and fluids (CORK model: (Holland & Powell, 1991)). Elemental fractionation during prograde garnet porphyroblast growth was considered for sample SB177 which has a garnet modal proportion of >1%, using the method according to Marmo *et al.* (2002). Solid solution models used for calculations of pseudosections for metapelitic samples were as follows:

Bio(TCC) for biotite (Tajčmanová, *et al.*, 2009), Chl(HP) for chlorite (Holland, *et al.*, 1998), St(HP) for staurolite (Holland & Powell, 1998), Ctd(HP) for chloritoid (Holland & Powell, 1998), Gt(WPH) for garnet (White, *et al.*, 2001), Ilm(WPH) for ilmenite, hCrd for cordierite, Mica(CHA1) for titanium-, magnesium- and iron-bearing white mica (Auzanneau, *et al.*, 2010; Coggon & Holland, 2002), melt(HP) for haplogranitic melts (Holland & Powell, 2001), Pl(h) for plagioclase feldspar and Kf for potassic feldspar.

For the garnet amphibolite (sample SB344), ferric iron content in amphiboles was calculated using the spreadsheet of Locock (2014) and was found to be negligible. The structural formulae and andradite content of garnet porphyroblasts were calculated according to the methods of Droop (1987). Given ilmenite is the only oxide observed in the sample and minimal ferric iron in the structural formula of garnet and amphibole, O_2 and Fe_2O_3 were excluded from calculations. Sample SB344 was modelled using the following solid solution models:

Gt(WPH) for garnet (White, *et al.*, 2001), GlTrTsPg for amphibole (Wei & Powell, 2003; White, *et al.*, 2003), Act(M) for actinolite, Ilm(WPH) for ilmenite, melt (HP) for haplogranitic melts (Holland & Powell, 2001), Pl(h) for plagioclase feldspar, Opx(HP) for orthopyroxene and Omph(GHP2) for

clinopyroxene (Diener & Powell, 2012; Green, et al., 2007).

4.3.3 Geochronology

4.3.3.1 In-situ SHRIMP U-Th-Pb monazite dating

Monazite grains were initially located and investigated with transmitted and reflected light optical microscopy for textural location and optical features. The internal structure of each grain was further characterised by back-scatterede electron (BSE) imaging using a JEOL FEG 7001F scanning electron microscope at the Monash Centre for Electron Microscopy, Monash University, Melbourne, Australia. Monazite grains were analysed in situ using the Sensitive High Resolution Ion Microprobe (SHRIMP II) at the Research School of Earth Sciences, the Australian National University, Canberra, Australia. To ensure monazites were dated in the correct textural and petrogenetic context, polished thin sections were cut and mounted in a 30mm diameter SHRIMP "megamounts." Large scale maps of the final mounts were constructed from reflected light photographic mosaics to target SHRIMP spot analysis. Monazite standard 44069, which has a U/Pb age of 425 Ma and U concentration of 2100 ppm (Aleinikoff, et al., 2006) was mounted in each epoxy mount with thin section segments. The analyses were performed using a \sim 5nA negative O primary beam and a selected spot size of 15 μ m. The secondary beam was energy filtered by ~70% to minimise scattered ions and interference on the common Pb isotope ²⁰⁴Pb and to reduce the Th peaks. Analyses of the unknowns measured 6 scans through the mass stations ²⁰³CeO₂, ²⁰⁴Pb. 204.1 (background), ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb, ²³²Th, ²³⁸U, ²⁴⁸Th and ²⁵⁴UO, regularly interspersed with analysis of the monazite standard. Data reduction was performed according to Williams (1998, and references therein), using SQUID 2 Excel macro of Ludwig (2009) and Excel macro Isoplot 3 (Ludwig, 2003). Age calculations used decay constants recommended by the IUGS Subcommission on Geochronology (as given in Steiger & Jäger, 1977) and corrections for common Pb were made using the measured ²⁰⁴Pb and the relevant common Pb compositions (Stacey & Kramers, 1975). Uncertainties of 1s are given for individual U-Pb analyses (ratios and ages), however, uncertainties in the calculated weighted mean ages are reported as 95% confidence limits and include the uncertainties in the standard calibrations where appropriate.

4.4 Petrography and mineral chemistry

Field work was conducted throughout the Sefwi Greenstone Belt and the adjacent Sunyani-Comoé and Kumasi-Afema domains to collect samples of the major metamorphic rocks and igneous suites within distinct litho-tectonic domains. Samples were selected based on the geographic location, structural context and informative mineral assemblages for geothermobarometry. Sample location, peak metamorphic assemblages and whole rock bulk composition for key samples are presented in Tables 4.1 and 4.2. Table 4.3 contains representative mineral compositions described in the text. Mineral abbreviations are after Whitney and Evans (2010).

Table 4.1. Peak metamorphic mineral assemblage and location of the investigated samples.

Sample	Latitude	Longitude	Lithology	Peak metamorphic assemblage
SB344	6.0693	-2.9028	Garnet amphibolite	Grt + Fprg + Pl + Ilm
SB389	7.4401	-2.1071	Pelitic paragneiss	Bt + Sil + Pl + Ms + Qz + L
SB398	7.4601	-2.1956	Migmatitic pelitic paragneiss	Grt + Bt + Ky + Pl + Ru + Qz + L + Gr
SB164	6.4427	-2.7891	Mafic amphibolite	Grt + Hbl + Cum + Pl + Ilm
SB177	7.4162	-2.1564	Mafic high-pressure amphibolite/ granulite	Grt + Cpx + Pl + Ilm + Ttn + Qz + L

Table 4.2. Whole rock bulk compositions (oxide wt %) of samples used for calculation of P-T pseudosections.

					wt	%				
Sample	SiO_2	TiO ₂	Al_2O_3	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	H_2O
SB344	52.653	1.628	13.498	15.209	0.219	5.245	8.725	2.638	0.185	exc.
SB389	62.815	0.739	19.155	6.497	0.077	2.817	1.891	2.778	3.231	exc.
SB398	65.401	0.701	18.623	7.012	0.079	2.667	1.270	1.594	2.652	exc.
SB164	49.578	1.419	17.312	13.384	0.197	6.746	10.478	0.765	0.122	exc.
SB177	50.131	1.073	17.401	10.065	0.158	3.484	14.884	2.602	0.203	exc.

Exc., excess

4.4.1 High-grade rocks

4.4.1.1 Sample SB398 (garnet-kyanite-rutile migmatitic paragneiss)

Sample SB398 is a locally migmatitic paragneiss from the western part of the Chiraa Domain. The outcrop comprised interbedded, gneissic metagreywackes and rare aluminous metapelitic layers. These rocks are intruded by centimetre- to metre-scale granitic dykes. Abundant and variable garnet and kyanite modal proportions in Al-rich layers indicate compositional heterogeneity in the protolith (Fig. 4.3a,b). The primary structures include a penetrative gneissosity (S_1) parallel to compositional layering defined predominantly by biotite alternating with quartz-feldspar layers. S_1 is folded around tight, symmetric metre-scale F_2 folds plunging moderately towards west-northwest to southwest. At the outcrop from which sample SB398 was derived (Fig. 4.3b), kyanite defines a prominent L_2 mineral stretching lineation, pitching ~30° to the west.

<i>1 uute</i> 4. <i>J</i> .	nepres	entativ	e chen	nical c	omposi	uons v	j miner	als in	metam	orpnic	assemoi	iages us	ea to c	constrau	1P-1c	onditic	ns ana	<i>P-1 pa</i>	uns fro	m pseuac	Sections			
Sample						SB398									SB389							SB34	4	
Mineral position	Gi	rim	Grt incl.	PII Grt core incl.	PI2 Grt mantle incl.	PI3 matrix	Ph1 Grt incl.	Ph2 matrix	Ms retro. rim on Ky	Bt1 Grt mantle incl.	Bt2 matrix	Git	Grt incl.	PI ps/morph of melt	PI matrix	Grt incl.	Bt1 incl. in Ms o/g Ky	Bt2 matrix	Ms matrix	Ms intergrown with Hem	Cort1 core	rim	Hprg matrix	PI matrix
SiO2	37.53	37.88	26.52	59.80	61.11	61.30	47.93	48.09	46.14	36.60	34.95	37.90	61.77	61.97	61.52	33.92	34.77	34.83	45.10	45.19	36.66	37.12	42.42	66.49
TiO2	0.02	b.d.l.	0.13	b.d.l.	0.01	b.d.l.	0.68	0.77	0.05	2.55	2.72	0.02	b.d.l.	b.d.1.	b.d.l.	1.63	3.13	2.58	0.59	1.42	0.15	0.03	0.78	0.00
Al2O3	21.01	20.92	22.07	25.69	24.62	24.50	35.02	34.34	36.34	19.78	19.21	20.90	24.50	24.48	24.30	18.86	17.31	18.30	32.75	33.41	20.77	20.83	13.91	22.47
Fetot as FeO	34.17	34.48	23.54	0.80	0.45	0.01	1.38	1.14	0.77	16.58	19.25	30.11	0.24	0.04	0.02	20.36	20.51	19.47	3.43	3.30	26.72	28.34	19.45	0.05
MnO	1.50	1.63	0.06	0.01	b.d.l.	0.04	0.05	b.d.l.	b.d.l.	0.03	0.05	5.67	b.d.l.	b.d.l.	0.01	0.31	0.22	0.18	0.07	b.d.l.	4.19	1.70	0.14	0.00
MgO	3.71	2.90	14.96	b.d.l.	0.02	b.d.l.	0.80	0.83	0.41	9.89	8.70	3.77	b.d.l.	0.01	b.d.l.	8.49	8.82	8.64	0.58	0.65	1.56	1.85	7.00	0.00
CaO	2.27	2.21	0.11	7.70	6.40	5.93	b.d.l.	b.d.l.	0.01	0.01	b.d.l.	1.85	6.15	6.08	6.08	0.04	0.07	b.d.l.	0.01	b.d.l.	9.48	9.81	10.77	3.39
Na2O	0.05	0.01	0.02	7.36	8.14	8.17	0.90	0.45	0.82	0.31	0.20	0.04	8.07	8.17	8.30	0.12	0.10	0.20	0.56	0.41	0.01	0.00	2.24	8.82
K20	b.d.l.	0.02	0.50	0.14	0.03	0.13	10.04	10.18	10.35	9.48	9.28	b.d.l.	0.09	0.09	0.22	9.47	9.78	9.85	10.89	10.81	0.00	0.00	0.31	0.07
Cr2O3	0.03	0.02	0.05	0.01	b.d.l.	b.d.l.	0.01	0.05	b.d.l.	0.07	0.01	0.01	b.d.l.	0.01	b.d.l.	0.06	0.07	0.06	0.08	0.04	0.00	0.00	0.00	0.00
BaO			b.d.l.				b.d.l.	0.21	0.31	0.20	0.16					0.10	0.31	0.27	0.58	0.43				
CI			0.14				b.d.l.	0.02	0.01	0.25	0.64					0.71	0.50	0.32	0.20	0.12				
F			0.01				b.d.1.	b.d.l.	b.d.l.	0.01	0.02					0.03	0.01	b.d.l.	b.d.1.	0.01				
Total	100.29	100.07	88.12	101.51	100.77	100.07	96.82	96.08	95.21	95.75	95.17	100.26	100.81	100.84	100.44	94.09	95.59	94.69	94.83	95.79	99.54	99.70	97.03	101.30
Oxygen	12	12	28	32	32	32	22	22	22	22	22	12	32	32	32	22	22	22	22	22	12	12	23	32
Si	6.00	6.07	5.51	10.54	10.80	10.88	6.26	6.32	6.14	5.47	5.37	6.05	10.88	10.91	10.89	5.34	5.39	5.40	6.18	6.11	5.92	5.96	6.45	11.50
Al ^{Iv}	0.00	0.00	2.49	5.34	5.13	5.12	1.74	1.68	1.86	2.53	2.63	0.00	5.09	5.08	5.07	2.66	2.61	2.60	1.82	1.89	0.08	0.04	1.55	4.58
AL	3.96	3.95	2.91				3.65	3.65	3.83	0.96	0.85	3.93				0.85	0.55	0.75	3.47	3.43	3.87	3.90	0.94	
Fe ³⁺	0.07	0.01		0.12	0.07	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.03	0.00	0.00	0.00	0.00	0.00	0.25	0.23	0.26	0.20		0.00
Ti	0.00	0.00	0.02	0.00	0.00	0.00	0.07	0.08	0.01	0.29	0.31	0.00	0.00	0.00	0.00	0.19	0.37	0.30	0.06	0.14	0.02	0.00	0.09	0.00
Cr	0.00	0.00	0.01				0.00	0.00	0.00	0.01	0.00	0.00					0.01	0.01	0.01	0.00	0.00	0.00	0.00	
Mg	0.89	0.69	4.63	0.00	0.00	0.00	0.16	0.16	0.08	2.21	1.99	0.90	0.00	0.00	0.00	1.99	2.04	2.00	0.12	0.13	0.38	0.44	1.59	0.00
Fe ²⁺	4.49	4.62	4.09	0.00	0.00	0.00	0.15	0.13	0.09	2.07	2.47	4.01	0.01	0.00	0.00	2.68	2.66	2.53	0.15	0.14	3.34	3.60	2.47	0.01
Mn^{2+}	0.20	0.22	0.01	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.77	0.00	0.00	0.00	0.04	0.03	0.02	0.01	0.00	0.57	0.23	0.02	0.00
Ca	0.39	0.38	0.02	1.45	1.21	1.13	0.00	0.00	b.d.l.	0.00	0.00	0.32	1.16	1.15	1.15	0.01	0.01	0.00	0.00	0.00	1.64	1.69	1.75	0.63
Na	0.02	0.00	0.01	2.52	2.79	2.81	0.23	0.11	0.21	0.09	0.06	0.01	2.76	2.79	2.85	0.04	0.03	0.06	0.15	0.11	0.00	0.00	0.66	2.96
K	0.00	0.00	0.13	0.03	0.01	0.03	1.67	1.71	1.76	1.81	1.82	0.00	0.02	0.02	0.05	1.90	1.93	1.95	1.90	1.86	0.00	0.00	0.58	0.01
$X_{\rm Prp}$	0.15	0.12										0.15									0.06	0.07		
X_{Alm}	0.76	0.78										0.67									0.58	0.62		
$X_{ m Sps}$	0.03	0.04										0.13									0.09	0.04		
X_{Grs}	0.06	0.06										0.05									0.09	0.10		
X_{Mg}	0.15	0.13	0.53							0.52	0.44	0.18				0.43	0.44	0.43			0.26	0.27	0.39	
$X_{ m Or}$				0.01	0.00	0.01							0.01	0.01	0.01									0.00
X_{Ab}				0.63	0.70	0.71							0.70	0.70	0.70									0.82
X_{An}				0.36	0.30	0.28							0.29	0.29	0.28									0.17
b.d.l. below detectio	n limit; incl.	inclusion; ps	√morph pseu	ıdomorph; o	/g overgrowii	ng																		



Fig. 4.3. Outcrop photos of SB398. a) Outcrop photo of SB398 showing refolded pelitic and psammitic layers and parallel biotite schistosity, with parallel to cross-cutting leucosomes and leucocratic veins; b) Aluminium-rich pelitic layer containing high modal proportions of kyanite and garnet with increasing biotite content proximal to stromatic leucomsones.

Petrographic relationships are used to define a succession of four distinct metamorphic assemblages, denoted A - D (Fig. 4.4). The first two assemblages are defined by two unique plurimineral aggregates in the core and mantle of subhedral poikilitic, garnet porphyroblasts (2–7 mm) (Fig. 4.4a). Garnet porphyroblast composition shows irregular zonation toward rim, consistent with back-scattered electron (BSE) images which indicate a thin, bright, irregular rim. Electron microprobe analysis (EMPA) transects of garnet porphyroblasts (Fig. 4.5a) display curved chemical profiles of Prp13-15, Alm75-77, Grs5-6 and Sps2-3, with decreasing pyrope (Prp ~10) and increasing spessartine (Sps~4) and almandine (Alm~78) on garnet rims indicating partial garnet resorption and re-equilibrium (Kohn & Spear, 2000; Spear, 1993). Garnet cores contain fine-grained, pluri-mineralic aggregates of anhedral plagioclase (Pl1), euhedral rutile, radial chlorite and accessory apatite, zircon and graphite (Fig 4.4b), with abundant, small (<10 µm) quartz inclusions. Chl has a #Mg (Mg/(Mg + Fe)) ranging from 0.29 to 0.53 with a negligible Ti content. Anhedral plagioclase inclusions (Pl1) in the core of garnet (Fig. 4.4a, b) have An content (=100 x (Ca/(Ca + Na)) of 32-40. Assemblage (A), defined as $Grt + Chl + Pl1 + Rt + Qz \pm Bt$, is consistent with greenschist to low-T amphibolite conditions. Pluri-mineral inclusions in the garnet mantle comprise euhedral biotite1 and phengite1 laths and coarse-grained, subhedral to euhedral plagioclase2, with prismatic to elongate, single grain rutile inclusions (Fig 4.4c, d). Biotite1 has #Mg of 50 - 52 and Ti^{vi} = 0.11-0.15 a.p.f.u. (based on 11 oxygen).



Fig. 4.4. Photomicrographs of sample SB398 (a, d, e) and BSE images (b, c, f). (a) Garnet porphyroblast with inferred garnet core-mantle boundary (red dashed line) defined by inclusion size and composition. Core inclusions comprise plagioclase1, chlorite and rutile, whilst mantle inclusions comprise biotite1, phengite1 and plagioclase2. Matrix minerals include biotite2, phengite2, plagioclase3 and deformed kyanite blades with quartz and biotite inclusion trails (right of image centre). Graphite is ubiquitous throughout the sample. (b) Assemblage (A) in garnet core comprising anhedral plagiolase1, radial chlorite laths and euhedral rutile. (c) Assemblage (B) in garnet mantle euhedral biotite1, phengite1 and coarse-grained plagioclase1. (d) Single grain inclusions of rutile in garnet mantle. (e) Subhedral biotite2 in the matrix in contact with monazite-bearing kyanite blade and matrix plagioclase3. (f) Retrograde muscovite and ilmenite in matrix, growing at the apex of a kyanite blade containing monazite.



Fig. 4.5. Chemical composition profiles of transects across garnet porphyroblasts in samples (a) SB398, (b) SB389, (c) SB344 and (d)SB177. Profiles show location of garnet cores (c) and rims (r).

White mica inclusions have high potassic white mica content ($Ms = 100 \times (K/K + Ca + Na)$ of Ms=88-92 and elevated TiO_2 content of 0.68–1.02 wt %. It has a composition of Si = 3.11–3.13 a.p.f.u, $Fe^{2+} = 0.07-0.08$ a.p.f.u. and Mg = 0.03-0.04 a.p.f.u. (11O). This composition indicates partial phengitic and Ti substitution by Si + $R^{2+} = Al^{vi} + Al^{iv}$ and $Al^{vi} + Si = Ti + Al^{iv}$, consistent with high- $P 2M_1$ phengites (Auzanneau, *et al.*, 2010; Cibin, *et al.*, 2008; Massonne & Schreyer, 1987), and is, therefore, denoted as phengite1 (Ph1). The euhedral shape and preservation of crystal faces of plagioclase2 (Fig. 4.4c) suggests crystallisation from a melt (Vernon & Collins, 1988). Plagioclase2 shows a small decrease in An (30 – 36). A second assemblage (B) is, therefore defined as Grt + Bt1 + Ph1 + Pl2 + Rt + Qz ± L (melt) associated high amphibolite facies metamorphism (e.g. Spear, 1993), potentially corresponding with high-P amphibolite facies conditions. The transition from metamorphic assemblage (A) in the garnet core to metamorphic assemblage (B) in the garnet mantle is interpreted as the product of two reactions. The first is the chlorite dehydration reaction (R1) corresponding to increasing pressures and temperatures:

$$Chl + Kfs + tk \rightarrow Ph1 + Bt + Qz + H_2O$$
(R1),

after Simpson *et al.* (2000), whereby *tk* is an additional Tschermak component $[Al_2(Fe, Mg)_{-1}Si_{-1}]$ exchanged with garnet (Coggon & Holland, 2002). The absence of Kfs inclusions in garnet may indicate its total consumption as a reactant, typical of average Al-rich, K-poor metapelites bulk compositions (Simpson, *et al.*, 2000; Spear, 1993). The shift from assemblage (A) to assemblage (B) indicates partial

melting of the rock and transition to suprasolidus conditions. The third assemblage (C) is defined by phases in the matrix and the appearance of kyanite. Garnet porphyroblasts are observed in contact with matrix biotite (Bt2) and subhedral, matrix plagioclase (Pl3) (An 27–36), displaying overlapping An content to Pl2 inclusions in garnet. Kyanite has two grain morphologies in the matrix indicating multiple stages of growth. Millimetric, irregular anhedral kyanite grains containing rare monazite grains are observed as irregular inclusions parallel to elongate matrix biotite (Bt2), with a composition of #Mg = 42-45 and Ti^{vi} = 0.12–0.17 a.p.f.u (based on 11 oxygen). Bt2 defines the penetrative metamorphic foliation in SB398. The more abundant form of kyanite is represented by millimetric to centimetric, near euhedral, poikilitic porphyroblasts. These contain inclusions of matrix biotite (Bt2), quartz, rutile, monazite and graphite aligned parallel or subparallel to the c-axis (Fig. 4.4a, e, f), thus kyanite is interpreted as syntectonic. Poikiloblastic kyanite is not observed in equilibrium contact with garnet porphyroblasts. Production of peritectic kyanite during anatexis is interpreted as the product of near total consumption of phengite, attributed to muscovite dehydration reaction 2 (R2) in suprasolidus conditions:

$$Ms (Ph2) + Pl + Qtz \rightarrow Ky + melt (L)$$
(R2),

consistent with reactions observed in low-temperature (<800 °C) anatectic pelites in the Greenland Caledonides (Kalsbeek, *et al.*, 2001).

Assemblage (C) comprises the phases Grt + Bt2 + Ky + Pl3 + Ph2 + Rt + Qz ± L. Remaining phengite in the matrix (Ph2) has a similar, although more variable, composition to phengite1 inclusions in the garnet with a composition of Si= 3.06-3.19 a.p.f.u, Al^{iv} = 0.81-0.83 a.p.f.u., Al^{vi} = 1.81-1.95a.p.f.u. and Mg = 0.04-0.08 a.p.f.u. (11O). Fe²⁺ in phengite2 displays a significant decrease from ~ 0.07a.p.f.u. to ~ 0.05 a.p.f.u.. Biotite dehydration in SB398 is limited by its coexistence with rutile, thus extending its stability under higher temperatures (Guidotti, 1984; Stevens, *et al.*, 1997). Furthermore, the inclusion of rutile grains in kyanite and as single grains in the matrix indicates that the transition from assemblage (B) to assemblage (C) remained at temperature and pressure conditions above the stability fields of ilmenite, at the amphibolite-granulite facies transition.

Early retrogression in sample SB398 is indicated by small muscovite overgrowths on kyanite rims (Fig. 4.4f), also overgrowing the biotite foliation associated with the back reaction of fluid produced during crystallisation of small volumes of melt, as defined by reaction 3:

$$Ky + Pl + H_2O \rightarrow Ms + Qz \tag{R3}.$$

Overgrowths of muscovite (Ms) have Ms = 89–91, and lower Si = 3.06-3.08 a.p.f.u. and Al^{vi} = 1.89-1.91 a.p.f.u. compared to Ph2 in the matrix. Low modal proportion (<1%) of late ilmenite in muscovite cleavage planes (Fig. 4.4f) indicates the rock crossed from the stability field of rutile into the stability field of ilmenite during retrogression. Garnet chemical profiles (Fig. 4.5a) show downward inflection in #Mg, coinciding with a decrease Prp and increased Alm and Sps content (Prp12, Alm78, Sps4) suggesting partial resorption along garnet rims (Kohn & Spear, 2000; Spear, 1993).

4.4.1.2 Sample SB389 (migmatitic biotite-sillimanite-garnet paragneiss)

Sample SB389 is from a migmatitic biotite-sillimanite paragneiss with relict garnet porphyroblasts, located nine kilometres northeast of Chiraa, in the Chiraa high-grade domain (Fig. 4.2). Sedimentary layering in outcrop is interpreted from irregular, alternating high- and low-modal proportions of micas and aluminosilicates in pelitic and psammitic beds, respectively (Fig. 4.6a). The metatexite has a penetrative biotite-fibrolitic sillimanite metamorphic foliation (S_1) alternating with quartz-plagioclase bands, featuring coarse fibrolitic sillimanite nodules, with abundant accessory apatite, and minor monazite, ilmenite, ferroceladonite-rich muscovite and hematite. Coarse-grained, foliation-parallel stromatic leucosomes observed in outcrop feed into centimetric quartz-plagioclase rich dykes (Fig. 4.6b) with biotite-rich selvedges forming on margins of quartz-plagioclase rich leucosomes (Fig. 4.6c). Field observations indicate limited melt segregation and minor melt migration. Centimetric, asymmetric, moderately inclined tight folds refold the S_1 gneissosity. A northeast-southwest striking, S_2 axial planar foliation, defined by biotite, dips 75° to the southeast. F_2 folds plunge moderately to the southwest.



Fig. 4.6. Field photos of biotite-sillimanite sample SB389. a) Outcrop showing strongly deformed meta-pelitic and metapsammitic layers, which sheared bedding parallel leucosomes and granitic dykes and leucosomes feeding into high-

Petrographic relationships and mineral chemical compositions are used to constrain three metamorphic assemblages (A-C). Garnet is preserved as relict, millimetric rounded-toglobular, anhedral porphyroblasts, with quartz, plagioclase (Pl1) and biotite (Bt1) inclusions (Fig. 4.7a), representing a modal proportion of less than 1%. Garnet often appears as disaggregated porphyroblasts, truncated along overprinting biotite foliations. Garnet composition is strongly zoned with a core composition of Prp12-14, Grs4-6, Sps12-17, Alm64-67 (Fig. 4.5b), with increasing Sps (16-20) and decreasing Prp (10-11) towards garnet rim indicate strong garnet resoption and element diffusion (Kohn & Spear, 2000; Spear, 1993). We note that the combined evidence of anhedral grain morphology and garnet resorption indicate that garnet compositions must be treated with caution. Anhedral biotite (Bt1) inclusions observed in garnet have a composition of #Mg = 43-45 and Ti^{vi} = 0.12–0.14 a.p.f.u. Plagioclase inclusions (Pl) in garnet have An(26–29), within the range observed for plagioclase in the matrix (An23-31). Muscovite observed in contact with garnet porphyroblasts has a composition of Ms = 93–97, Si = 3.04-3.12 a.p.f.u. and Al^{vi} = 1.68-1.80 a.p.f.u., with Fe²⁺ = 0.02-0.16a.p.f.u. and $Fe^{3+} = 0.08-0.16$ a.p.f.u., calculated using charge balance between charge ideal nd charge real, indicating partial ferric iron-rich celadonite substitution, according to Tischendorf et al. (2004). The composition is comparable to muscovite inclusions in subhedral matrix plagioclase. Metamorphic assemblage (A) is defined as $Grt + Bt1 + Pl + Qz \pm Ms$, however, we note that microstructural evidence for the prograde assemblage associated with garnet porphyroblasts in this sample is often ambiguous due poor preservation.

Relict kyanite porphyroblasts occur in the matrix with extensively corroded rims and never in equilibrium contact with garnet. Figures 4.7a and 4.7c show a kyanite porphyroblast with a subrounded quartz core and an interstitial cuspate film of plagioclase (An 29–31) pseudomorphing melt, suggesting kyanite was forming in suprasolidus conditions and respresents a peritic phase (Holness & Sawyer, 2008). Kyanite is interpreted as the product of muscovite dehydration melting during reaction 2 (R2) as the rock crossed the solidus. Potassium feldspar was observed only as rare, round inclusions in plagioclase in thin section, suggesting limited production during reaction 2, total consumption during subsequent retrograde reactions or extraction during melt segregation and transportation. Reaction 2 occurs at the transition from assemblage (A) to assemblage (B) Grt + Bt2 + Ky + Pl + Qtz + L at suprasolidus, high-pressure amphibolite facies conditions.



Fig. 4.7. Photomicrographs of migmatitic metapelite sample SB389. (a) Early, rounded, anhedral garnet porphyroblast in contact with muscovite, quartz and plagioclase. Aligned biotite defines penetrative metamorphic fabric. (b) Isolated corroded kyanite porphyroblast with thick overgrowth of muscovite with the square indicating inset (c). (c) A cuspate film of plagioclase pseudomorphing melt between corroded kyanite and rounded, anhedral quartz. Biotite2 visible as inclusion in muscovite2. (d) BSE image of randomly aligned, fine grained, fibrolitic sillimanite in sillimanite nodule. (e) Fibrolitic sillimanite is observed in two morphologies; the first, as coarse-grained knots suggesting it is a retrograde phase, pseudomorphing pre-existing porphyroblasts; the second textural location of fibrolitic sillimanite is sub-parallel to the S, foliation. Square in sillimanite nodule shows location of (d), not to scale.

Kyanite porphyroblasts are embayed in thick muscovite overgrowths containing small round inclusions of quartz and biotite (#Mg = 43-44; Ti^{vi} = 0.17-0.18 a.p.f.u.). Muscovite composition is indistinguishable from that of muscovite in contact with garnet. The relationship between kyanite

and muscovite overgrowths is interpreted the product of two possible reactions: first, as a rapid metasomatic back reaction between kyanite and water released during *in situ* melt crystallization (R3) occurring as the rock returned below the solidus, and second, the following closed-system sillimanite-forming, decompression, chain reaction (R4):

$$Ky + Bt + Qz \rightarrow Ms \rightarrow Sil + Bt + Qz$$
 (R4),

involving mica as a catalyst (Foster, 1991; Gervais & Hynes, 2013). This reaction produces distinct replacement of kyanite by muscovite with Bt-free mantles and sillimanite growth after biotite with Ms-free mantles forming fine-grained, fibrolitic sillimanite nodules, as described by Foster (1991) for high-grade pelites. R4 and R5 are responsible for the formation of the dominant matrix assemblage (C), Sil + Bt + Pl + Qz \pm Ms \pm Grt. Elongate, fibrolitic sillimanite is also observed sub-parallel to the subhedral, elongate, highly-pleochroic biotite (#Mg 42-45 and Ti^{vi}= 0.10–0.15 a.p.f.u) and muscovite foliation suggesting metamorphism is concurrent with deformation (Fig. 4.7d,e), representing the biotite-producing end of R5. Matrix plagioclase has An 27–29. This assemblage is often seen over-printing relict kyanite grains and dark brown, highly corroded biotite grains with round quartz inclusions, representing the early biotite consumption of R5. Despite differences in morphology and microstructural context, the two types of biotite have indistinguishable chemical composition.

Late oxidation of the rock is indicated by an overprint of matrix biotite by an intergrowth of muscovite (Ms2) and hematite. The oxidation of biotite and aluminosilicate produces muscovite and hematite, modelled by the following balanced chemical equation: $4Bt + 4Sil + O_2 = 4Ms + 6Hem + 4Qz$. Ms2 composition displays lower Si content than from Ms1, with a composition of Ms = 93–96, Si = 3.03-3.05 a.p.f.u. and Al^{vi} = 1.71-1.79 a.p.f.u..

4.4.1.3 Sample SB344 (garnet-bearing amphibolite)

Sample SB344 is sourced from an outcrop located 5 km south of the Ketesso shear zone, 40km south of Juaboso (Fig. 4.2). The location of this garnet amphibolite is interpreted as the southernmost extent of the amphibolite-facies Kukuom-Juaboso domain based on its metamorphic grade and structural information. Exposures of chlorite-actinolite bearing, weakly-foliated metabasalts and unmetamorphosed granodiorite plutons are found less than 5 km to the southeast of the garnet amphibolite sample (SB344), therefore, we infer a significant lateral metamorphic break lies to its south and southeast, likely coinciding with a poorly exposed tectonic contact. The penetrative metamorphic foliation in this sample is defined by aligned subhedral to euhedral, elongate amphibole grains and minor plagioclase and quartz-rich bands, wrapping around millimetric garnet porphyroblasts. Strain shadows of garnets primarily consist of quartz. Accessory apatite is visible throughout the sample.

Garnet porphyroblasts are subhedral to euhedral, 1 - 3 mm in diameter with aligned inclusions of elongate quartz and magnesio-ilmenite, which preserve a relict foliation (S₁) oblique to the penetrative amphibole and plagioclase foliation of the matrix, denoted as S_2 (Fig. 4.8a). The obliquity between the two foliations suggests garnet growth is inter-tectonic (Fig. 4.8b). Measured garnet composition shows slight zoning from core to rim with Prp6 \rightarrow Prp7, Grs26 \rightarrow Grs27, Sps10 \rightarrow Sps4 and Alm58 \rightarrow Alm62 (Fig. 4.5c). Plagioclase in the matrix has a composition of An 15–21, with one transect revealing a higher An content (An24-30). Matrix amphibole is classified as ferropargasite (Fprg) with #Mg = 39-41, Ca/Na = 2.73-3.11, (Ca+Na)_B = 1.79-1.89 a.p.f.u., (Na+K)_A = 0.56-0.68 a.p.f.u. and Ti^{vi} = 0.09 a.p.f.u.. The structural formula of ferropargasite was calculated according to the IMA 2012 recommendation, with the use of the excel spreadsheet compiled by Locock (2014). Ilmenite in the matrix is commonly aligned with the foliation and as inclusions in ferropargasite. Minor epidote is observed in the matrix. Assemblage (A) (Grt + Pl + Fprg + Ilm + Ep + Qz) represents the peak assemblage, corresponding with amphibolite facies conditions. Weak retrogression and secondary hydration is indicated by chlorite overgrowth, localised along foliation planes and thin replacement rims on garnet grain boundaries and internal fractures. Incomplete replacement of ilmenite by titanite is confined to S_2 foliation planes, with rounded ilmenite inclusions observed in foliation parallel titanite grains, suggesting a transition from the stability field of ilmenite to titanite (Fig. 4.8b). These phases define assemblage (B) Grt + Pl + Fprg + Ep + Chl + Ttn, indicative of high-P greenschist facies conditions.

4.4.1.4 Sample SB164 (garnet-bearing amphibolite)

Sample SB164 comes from the western margin of the amphibolite facies Kukuom-Juaboso domain in the Sefwi greenstone belt, about 10 km north of the township of Juaboso. The domain is separated from the chlorite-white mica schists of the Sunyani-Comoé metasedimentary province by a north-northeast striking shear zone (Fig. 4.2). The rock contains a high modal proportion (~30%) of subhedral garnet porphyroblasts (2–7 mm) and a penetrative foliation.



Fig. 4.8. Photomicrographs of metamorphosed mafic rocks. (a) Subhedral garnet porphyroblasts with inclusion trails of ilmenite and quartz, defining early metamorphic foliation (S_1) , hosted in a matrix of aligned, elongate ferropargasite, plagioclase and coarse-grained ilmenite. Strain shadows of garnet are filled with quartz. (b) Retrograde chlorite in along fractures in garnet porphyroblasts and overgrowth of rounded titanite around ilmenite along $S_{2,3}$ foliation plane. (c) Polymineralic inclusions of cummingtonite, magnesio-hornblende, quartz, plagioclase and ilmenite in garnet porphyroblasts in amphibolite sample SB164. (d) Early foliation preserved in garnet porphyroblasts at an oblique angle to the penetrative metamorphic foliation (S_2) in the matrix defined by elongate hornblende and plagioclase, seen wrapping around garnet porphyroblasts. (e) Pluri-mineralic inclusions in garnet of euhedral plagioclase1, subhedral hornblende1 and ilmenite and isolated inclusions of anhedral clinopyroxene1. (f) Peritectic garnet and clinopyroxene2 hosted in plagioclase-rich leucosome (Pl2) with titanite, ilmenite and clinopyroxene along leucosomes selvedges. A thin rim of hornblende2 and actinolite on the garnet porphyroblast comprise the early retrograde assemblage.

Unzoned, poikiloblastic garnet porphyroblasts (Prp15-18, Sps2-3, Grs14-16, Alm63-67) preserve aligned inclusion trails of magnesio-hornblende (Mhb), cummingtonite (Cum), plagioclase and ilmenite (Fig. 4.8c). The first assemblage (A) is represented by Grt + Mhb + Cum + Pl + Ilm + Qz. Magnesio-hornblende inclusions in garnet display the same composition as matrix magnesiohornblende, which represents ~70% modal proportion of matrix minerals. It has a composition of #Mg = 0.53-0.57, Ca/Na = 4.48-6.01, (Ca+Na)_B = 1.34-1.81 a.p.f.u. and (Na+K)_A = 0.27-0.39 a.p.f.u. Cummingtonite inclusions have #Mg 0.57-0.61, Ca/Na = 2.86-4.59, (Ca+Na)_B = 0.11-0.53 a.p.f.u. and $(Na + K)_A = 0.00-0.04 a.p.f.u.$ Both plagioclase inclusions and matrix plagioclase have a composition within the range of An88–93. Rare chlorite grains in garnet have #Mg = 43–46, otherwise absent from the matrix. The foliated matrix of SB164 (Fig. 4.8d) comprises the same assemblage as the garnet inclusions trails with the exception of cummingtonite and chlorite, which are not observed in the matrix. Epidote is only observed in the matrix. Assemblage (B) is defined as Grt + Mhb + Pl + Ilm + Ep + Qz. Both assemblages indicate epidote-amphibolite to amphibolite facies conditions. Whilst the obliquity between garnet inclusion trails and the foliated matrix suggest the garnet porphyroblasts are intertectonic (Passchier & Trouw, 2005), the mineral assemblages and mineral chemical compositions are too similar to distinguish multiple events without additional geochronological constraints (Vernon, et al., 2008).

4.4.1.5 Sample SB177 (garnet-clinopyroxene bearing migmatitic gneiss)

Sample SB177 is from a raft of migmatitic, calcic metabasalt surrounded by extensive granite outcrop 5 km east of Chiraa (Fig. 4.2). The outcrop SB177 contains large blocks of megacrystic metamorphic rocks preserving high-grade garnet-clinopyroxene-bearing metamorphic assemblages. The tectonic fabric strikes east-west, dipping steeply to the south with a gently west-plunging mineral stretching lineation. Leucosomes occur as discrete, disconnected patches, often pooling in fold hinges suggesting syn-tectonic partial melting and minimal melt segregation and loss. Peritectic garnet porphyroblasts are subhedral with irregular, embayed grain boundaries, often occurring in contact with peritectic clinopyroxene porphyroblasts. Garnets are heavily fractured.

Three metamorphic assemblages have been deduced from petrographic relationships and mineral compositions. Chemical profiles of garnets show strong zonation in grossular and almandine content (Fig. 4.5d). Changes in garnet composition from core (Grt1) to rim (Grt2) are Grs44–46 \rightarrow Grs 39–43, Alm49–50 \rightarrow Alm51–54, Prp3–4 \rightarrow Prp 5–6 and Sps1–2 \rightarrow Sps0–1 (Fig. 4.5d). Pluri-

mineral aggregates and singular inclusions of titanite and ilmenite, and plagioclase and hornblende constrain an early, amphibolite facies metamorphic assemblage (Fig. 4.8e), defined by (A) = Grt1 + Hbl1 + Pl1 + Ilm + Ttn + Mag \pm Cpx \pm Ep. Pl1 has An41–43 and Hbl1 has #Mg = 28–34, Ca/Na = 4–6, (Ca+Na)_B = 1.89–1.93 a.p.f.u. and (Na+K)_A = 0.51–0.61 a.p.f.u. Clinopyroxene in all textural locations as classified as Ca-Mg-Mg (Quad) clinopyroxenes. Clinopyroxene inclusions in garnet display an intermediate diopside-hedenburgite composition with #Mg = 0.44–0.49, Ca/(Ca + Na) = 0.97–0.99, high wollastonite content of Wo(100 *Ca/(Ca+Mg+Fe)) = 47–49, enstatite (En) (100*Mg/(Mg+Ca+Fe)) = 22–26 and ferrosillite (Fs) (100*Fe_T/(Fe_T+Ca+Mg)) = 24–28, after Morimoto *et al.* (1988).

Metamorphic assemblage (B) records the transition to suprasolidus conditions under the upper amphibolite to granulite conditions. It comprises Grt2 + Cpx + Pl2 + Qz + Ttn + Ilm \pm Czo + melt (L) (Fig. 4.8f). Garnet and clinopyroxene porphyroblasts are hosted in granoblastic plagioclase-rich leucosomes, where grain boundaries have visible 120° angles. Plagioclase in leucosomes is classified as Pl2, which has higher An content (62-71) than Pl1 inclusions in garnet. Clinopyroxene porphyroblasts hosted in leucocratic domains have #Mg 50–55, Ca/(Ca + Na) = 0.97–0.99, Wo= 49–51, En = 26–27 and Fs = 22–25. Clinopyroxene selvedges grow in equilibrium with both titanite and ilmenite on leucosomes margins (Fig. 4.8f). Early garnet and clinopyroxene growth are interpreted as the product of progressive dehydration of hornblende defined by the following reaction (R5), typical of transitional amphibolite metamorphism:

$$Hbl1 + Qz \rightarrow Grt1 + Cpx1 + Pl1 + H_2O$$
(R5).

The bulk rock composition has a low Na/Ca ratio, thus prolonging the hornblende-out reaction. Anatexis and peritectic growth of garnet and clinopyroxene occur during continued hornblende dehydration (R6) as the rock crosses the solidus:

$$Hbl1 + Pl1 + Qz = Grt2 + Cpx2 + L (or V)$$
(R6)

The peak metamorphic assemblage corresponds with key high-pressure granulitic associations in mafic rocks, forming at the boundary between amphibolite/high-pressure granulite metamorphic conditions (700–850 °C, 10–14 kbar) (O'Brien & Rötzler, 2003). The absence of orthopyroxene in garnet-clinopyroxene-bearing mafic granulites is indicative of partial melting occurring above 10 kbar (Pattison, 2003). Growth of retrograde hornblende (Hbl2) on garnet rims indicates the back reaction (R7) between peritectic minerals and melt, common in rocks with minimal melt extraction:

Hbl2 has #Mg = 34-41, Ca/Na = 5-6, (Ca+Na)_B = 1.95-1.99 a.p.f.u. and (Na+K)_A = 0.48-0.53 a.p.f.u. Continued hydration of garnet, and hornblende results in the final retrograde assemblage (C) of marked by the appearance of actinolite, chlorite, epidote, albite and calcite on late fracture networks and grain boundaries of porphyroblasts (Fig. 4.8f) and is defined as Hbl2 + Pl + Act + Ep + Chl + Ab. The secondary retrograde reaction (R8) is typical of greenschist facies metamorphism:

 $Grt2 + Hbl2 + Pl + H_20 \rightarrow Act + Chl + Ab + Ep + Calcite$ (R8).

4.4.2 H₂O-saturated melting

Partial melting of mafic protoliths occurs extensively along the Hwidiem shear zone and in the core of the Kukuom-Juaboso amphibolite domain. Peritectic hydrous minerals, predominantly comprising hornblende were commonly entrained in plagioclase-rich and quartz-rich leucosomes. Migmatitic amphibolites in the Hwidiem shear zone (e.g. SB089) host foliation-parallel leucosomes comprising subhedral hornblende and plagioclase with round quartz inclusions, with accessory epidote, apatite and rare euhedral titanite. A minor greenschist facies overprint comprises epidote, rutile and chlorite. Leucocratic domains in migmatitic amphibolite gneiss SB435 are observed with both diffuse boundaries or with hornblende selvedges (Fig. 4.9a). The leucosomes assemblage consists of euhedral plagioclase, centimetric, poikilitic hornblende prisms (Fig. 4.9b) and minor quartz and k-feldspar with accessory allanite and magnetite. Extensive alteration of the assemblage is marked by the presence of sericite, chlorite, epidote and albite. In the core of the Kukuom-Juaboso domain, an exposure of migmatitic gneissic metadacite (SB016) with a gneissosity defined by aligned amphibole alternating with quartzofeldspathic bands.



Fig. 4.9. (a) Leucosome with hornblende selvedges in migmatitic amphibolite gneiss located along the Hwidiem shear

4.5 Results

4.5.1 P-T conditions and P-T paths

4.5.1.1 Sample SB398 (garnet – kyanite - bearing paragneiss)

P-T pseudosections for paragneiss sample SB398 (Fig. 4.10) show stability fields of observed assemblages and mineral stability fields derived from measured mineral compositions. Observed pluri-mineralic inclusions of anhedral plagioclase, radial chlorite and euhedral rutile in garnet porphyroblasts cores potentially partially record an early prograde assemblage (A) at temperatures below the calculated Chl-out—Grt-in line at 590 °C (Fig. 4.10a) and pressures above 10 kbar as indicated by the inclusion of rutile. Subsolidus high-pressure amphibolite conditions of 9.9-11.5 kbar, 590-640 °C are deduced from measured garnet composition (Prp13-15, Alm75-77, Grs5-6) and biotite inclusion composition (Bt1 Mg# = 50-52), corresponding to the calculated stability field of Grt + Bt + Pg + Ms + Rt + Qz. This assemblage partially resembles that of assemblage (B), $Grt + Ph1 + Bt1 + Pl2 + Rt + Qz \pm$ L. The inclusions of euhedral plagioclase crystals in the mantle of garnet porphyroblasts potentially suggesting the assemblage is metastable as it first crossed the solidus. Paragonite (Pg) was not preserved as inclusions in garnet, nor in the matrix, potenitially indicating a preservation issue. The transition from assemblage (B) to assemblage (C) records heating concomitant with < 2 kbar of exhumation indicating suprasolidus upper amphibolite facies conditions of 690 °C at 9.5 kbar within the stability field of kyanite. The calculated stability field associated with these conditions has the assemblage Grt + $Bt + Ky + Pl + Ms + L \pm Rt$, corresponding well with the observed assemblage (C) Grt + Bt2 + Ky + Pl2+ Qz + Rt \pm Ms \pm L. At the calculated suprasolidus conditions, calculated plagioclase composition is consistent with that of measured euhedral garnet inclusions of plagioclase (Pl2) and matrix plagioclase (Pl3) have anorthite content (30-36), corresponding also to 4-6 % volume of kyanite (Fig. 4.10c). Calculated conditions and observations are consistent with estimations for partial melting of similar pelitic lithologies in the Caledonides by Kalsbeek et al. (2001) of ~ 750 °C and 10-12 kbar. Monazite is observed as inclusions in matrix biotite and in kyanite porphyroblasts (Fig. 4.4e, f) and is interpreted to have crystallised during peak temperature metamorphism.

The retrograde path is more poorly constrained. Phases associated with retrogression are limited in modal proportion, evident in small muscovite overgrowths on kyanite rim, the appearance of ilmenite after rutile in the matrix, often in assocaition with muscovite and decrease in the #Mg of


Fig. 4.10. MnNCKFMASHTi P-T pseudosections for phase equilibria for metapelite sample SB398. (a) Calculated stable assemblages. (b) Calculated garnet isopleths for pyrope (dotted lines and values in circles) and almandine (dashed lines and values in squares). Colour spectrum represents garnet volume. (c) The colour spectrum shows modal proportions of aluminosilicates with anorthite content of plagioclase in dashed lines with values in circles. (d) Isopleths of Bt #Mg in dashed line with value in squares.

Bt2, however, this likely indicates a cation exchange between biotite and garnet during garnet resorption and diffusional reequilibration. Matrix biotite in contact with garnet rims records a decrease in Mg# (#Mg 42 – 45) (Fig. 4.10d). The observed phases and compositions correspond to a calculated stability field with the assemblage Bt + Chl + Pl + Pg + Grt + Ilm and metamorphic conditions of ~ 6 kbar and 580–590 °C. The absence of retrograde sillimanite and staurolite has two possible explanations, the first of which suggests that concurrent cooling and decompression held the rocks at conditions above that of the stability field of sillimanite, preserving the kyanite porphyroblasts. The second explanation suggests problematic crystallisation due to decreasing temperature or fluid-absent conditions during retrogression.

Overall, *P-T* modelling reveals a clockwise *P-T* path for sample SB398. Garnet porphyroblasts record burial along a cool geothermal gradient (~15° C/km) crystallising under high-pressure amphibolite conditions near the amphibolite-granulite boundary (Vernon & Clarke, 2008). Following peak pressure, sample SB398 records a near isobaric increase in temperature to amphibolite facies suprasolidus conditions crossing geotherms and recording a transient apparent geothermal gradient of 20–25 °C/km. The retrograde path records a 3 kbar decrease in pressure and a drop of 100° C.

4.5.5.2 Sample SB389 (biotite-sillimanite gneiss)

P-T pseudosections for sample SB389 are presented in Figure 4.11. Measured compositions of relict garnet porphyroblasts (Grs 4-6, Sps12-17 and Prp12-15) correspond with calculated stability fields with assemblages Bt + Grt + Pl +Ms + Ilm + Qz and Bt + Grt + Pl +Ms + Qz (Fig. 4.11b) consistent with the observed petrological assemblage (A) Grt + Bt1 + Ms1 + Pl + Qz. Values of Ti^{iv} (0.12 – 0.14 a.p.f.u.) in biotite are consistent with the stability filed of Bt + Grt + Pl + Ms + Ilm + Qz. The absence of ilmenite suggests ilmenite consumption may have gone to completion or been replaced during subsequent alteration. Garnet prophyoblasts have been strongly resorbed, therefore the calculated stability field of garnet 5.5–9 kbar and 550–640 °C, associated with an apparent geothermal gradient of ~20 °C/km, are regarded with caution. Modelled garnet modal proportions indicate garnet represents less than 2% of phases present under these pressure and temperature conditions, consistent with petrological observations (Fig. 4.11b).

The prograde path is constrained by the preservation of highly corroded relict kyanite porphyroblasts with plagioclase melt rims. Modelling indicates modal proportions of kyanite under 5 % (Fig. 4.11c). The anorthite composition of plagioclase (An 29 - 31) melt rims around kyanite indicates peak metamorphic conditions of 8.5 kbar and 690 °C, coinciding with modelled phase assemblage of Bt + Pl + Grt + Ky + L ± Ms, matching assemblage (B). The transition to observed assemblage (B) Grt + Bt2 + Ky + Pl + Qtz + L is the product of reaction 4 and occurs under suprasolidus conditions at the amphibolite-granulite facies boundary. The overgrowth of muscovite on relict, resorbed kyanite grains indicates the rock returned below the wet solidus prior to a decrease in pressure into the stability



Fig. 4.11. MnNCKFMASHTi P-T pseudosections for phase equilibria for metapelite sample SB389. (a) Calculated stable assemblages. (b) Calculated garnet isopleths for pyrope (dotted lines and values in circles) and grossular (dashed lines and values in squares). Colour spectrum represents garnet volume. (c) The colour spectrum shows modal proportions of aluminosilicates with anorthite content of plagioclase in dashed lines with values in circles. (d) Isopleths of Bt #Mg in dashed line with value in squares presented with Ti content of biotite (grey dotted line and values in circles).

fields of sillimanite. The retrograde path is constrained by the pseudomorphing of kyanite porphyroblasts by sillimanite. Modelling indicates sillimanite modal proportions reach 5% as pressure drops below 7 kbar and 675 °C. Measured matrix biotite compositions (43–45) (Fig. 4.11d), matrix plagioclase anorthite content (27–29) and the appearance of sillimanite corresponds with the modelled phase assemblage of Bt + Pl +Ms + Grt + Sill + Qz (Fig. 4.11c). This is consistent with assemblage (C):

Sill + Pl + Bt + Ms + Qz \pm Grt. The absence of cordierite suggests either that pressure did not fall below 4 kbar during the initial stages of retrogression or alternatively that there was limited free water during exhumation prohibiting the crystallisation of retrograde phases. The absence of staurolite similarly suggests temperatures and pressures remained above ~5 kbar and > 590 °C, however, the evidence is inconclusive.

4.5.5.3 Sample SB344 – Garnet amphibolite

P-T pseudosections for garnet amphibolite sample SB344 (Fig. 4.12) shows calculated stable assemblages, observed assemblages and measured mineral compositions. Calculated mineral



Fig. 4.12. MnNCFKMASHTi P-T pseudosections for phase equilibria for metabasalt sample SB344.
(a) Calculated stable assemblages. (b) Calculated garnet isopleths for pyrope (dotted lines and values 155

in circles) and grossular (dashed lines and values in squares). Colour spectrum represents garnet volume. (c) Calculated magnesium number of amphibole values indicated by dashed lines and values in squares and anorthite content of plagioclase in dotted lines with values in hexagons.

structural formulae for garnet and amphibole showed negligible ferric iron content, therefore, ferric iron was not incorporated in pseudosection calculations and epidote is represented by zoisite in the pseudosections.

Peak pressure and temperature conditions of 10.0 - 10.5 kbar and 580 - 640 °C are deduced from grossular and pyrope content of garnet porphyroblasts, overlapping with a calculated assemblage of Grt + Amp + Pl + Ilm + Qz, consistent with the observed assemblage (A). These metamorphic conditions correspond with a slightly cool apparent geothermal gradient of ~17 °C km⁻¹ in the upper amphibolite facies. Following the formation of assemblage (A), the rock crossed hydration reaction 1, into the stability domain of chlorite, as indicated by the partial hydration and growth of chlorite in the fracture network of garnet and along the foliation plane. A further decrease in pressure and temperature crossing out of the stability domain for ilmenite and into that of titanite during reaction 2. The measured #Mg of amphibole and anorthite content of matrix plagioclase constrain greenschist facies conditions of 6.5–9 kbar and 450–500 °C, corresponding with a calculated assemblage of Chl + Amp + Pl + Grt + Zo + Ttn + Qtz and the observed matrix and retrograde assemblage.

4.5.5.4 Samples SB164 (garnet amphibolite) and SB177 (garnet-clinopyroxene bearing migmatitic gneiss)

Whilst it is mineralogically similar to sample SB344, attempts to model pressure and temperature condition for garnet amphibolite SB164 using Perple_X were unsuccessful. All pseudosections calculated for the bulk composition of SB164 yielded unresolvable discrepancies between observed mineral assemblages and measured and calculated mineral compositions. Pseudosections calculated for the garnet-clinopyroxene-bearing mafic migmatite (SB177) indicated unrealistic clinopyroxene stability down to greenschist facies. This is interpreted as the result of elevated CaO content in the bulk composition. Whole rock CaO wt% was corrected for calcite content, however, isopleths derived from the pseudosections with adjusted whole-rock CaO significantly underestimated Ca content relative to the measured Ca composition in the Ca-bearing phases (Grt, Cpx, Pl). Whilst we are confident the petrological and textural evidence in SB177 indicates high grade metamorphism of a mafic to intermediate metavolcanic rock, thermodynamic calculations of *P-T* for SB177 are deemed unreliable.

4.5.2 Geochronology

4.5.2.1 Sample SB398 (garnet-kyanite- bearing paragneiss)

Eighteen U-Th-Pb SHRIMP analyses were performed on 14 individual monazite grains from sample SB398 (mounts Z6388 and Z6391). All monazites are unzoned and hosted in biotite and kyanite porphyroblasts or as isolated grains in the matrix. Monazites in this sample are subhedral to anhedral, often featuring prominent elongation. *R*are inclusions of quartz occur in larger grains. Some monazites



Fig. 4.13. Results of U-Pb dating of samples SB398 and SB389. (a, b, d) Back-scattered electron images of monazites showing textural context. (a) Deformed monazite in kyanite porphyroblast in sample SB398. (b) SHRIMP analysis points

	•			•															
Grain Spot	Location	% ²⁰⁶ Pb _c	n U	ppm Th	²³² Th / ²³⁸ U	%∓	(1) ppm $^{206}Pb^*$	(1 ²⁰⁶ Pb) / ²³⁸ U 5e	(1) ²⁰⁷ Pb/ ²⁰ Age	'nb	Dis- cor- dant	$(1) \\ {}^{207} { m Pb}^{*} \\ /^{206} { m Pb}^{*}$	%∓	(1) $^{207} Pb^*$ $/^{235} U$	≁%	(1) $^{206}Pb^{*}/^{238}U$	+%	err corr
SB398 - I	Mount Z6388																		
1.1	Elongate in matrix	0.01	478	2452	5.30	0.71	152	2030	±25	2077	十7	+3	0.12845	0.42	6.56	1.5	0.370	1.5	0.961
1.2	Elongate in matrix	0.02	441	2346	5.50	0.71	144	2081	±25	2073	± 3	0-	0.12817	0.16	6.73	1.4	0.381	1.4	0.994
10.1	In matrix	0.02	290	1727	6.15	0.72	96	2100	±24	2073	+8	-1	0.12818	0.45	6.80	1.4	0.385	1.4	0.949
11.1	On Ky rim	0.13	376	2238	6.14	1.29	137	2278	±37	2067	LŦ	-12	0.12772	0.43	7.46	2.0	0.424	1.9	0.976
12.1	In matrix	0.01	358	2077	6.00	1.32	108	1949	±28	2080	士5	L+	0.12870	0.30	6.26	1.7	0.353	1.6	0.984
14.1	In matrix Bt	0.02	281	2033	7.47	0.71	71	1667	± 17	2068	±12	+22	0.12778	0.69	5.20	1.3	0.295	1.1	0.853
17.1	Deformed parallel to Ky pornhvroblast	0.01	259	1670	6 66	0.50	73	1874	+14	2073	۲ +	+14	0 12814	0.17	5 78	6.0	0 327	6.0	0.981
17.2	Deformed parallel to Ky				2		2				ì	-							
	porphyroblast	0.03	370	2320	6.47	0.74	101	1773	± 13	2076	1	+17	0.12837	0.52	5.60	1.0	0.317	0.9	0.853
17.3	Deformed parallel to Ky porphyroblast	1.08	345	1944	5.82	1.52	109	2025	±33	2068	4	+2	0.12781	0.54	6.50	2.0	0.369	1.9	0.962
SB398 - N	fount Z6391																		
1.1	Elongate in matrix Bt	0.01	861	4098	4.92	0.75	304	2218	±45	2074	± 11	8	0.12822	0.62	7.26	2.5	0.411	2.4	0.968
1.2	Elongate in matrix Bt	0.02	607	3394	5.78	0.91	200	2092	±46	2075	± 12	-1	0.12834	0.65	6.78	2.6	0.383	2.6	0.969
2.1	Elongate in matrix	0.00	969	3025	4.49	0.69	243	2202	±49	2071	± 12	L-	0.12802	0.69	7.19	2.7	0.407	2.6	0.967
3.1																			
	Stretched and elongate in matrix	0.02	911	4273	4.84	1.59	339	2320	±97	2080	±7	-14	0.12867	0.41	7.68	5.0	0.433	5.0	0.997
5.1	In corroded matrix Bt	0.01	640	3528	5.69	1.79	204	2035	±42	2071	± 3	+2	0.12799	0.17	6.55	2.4	0.371	2.4	0.997
6.1	Elongate in matrix Bt, truncated by																		
	Ky	0.01	594	3351	5.83	0.86	189	2033	±37	2078	τŦ	+3	0.12855	0.39	6.57	2.1	0.371	2.1	0.984
7.1	Elongate in matrix Bt	0.01	458	2720	6.13	0.92	142	1983	±37	2075	±5	+5	0.12830	0.29	6.37	2.2	0.360	2.1	0.991
8.1	In Ky porphyroblast	0.00	009	3411	5.87	0.86	204	2150	± 44	2082	土5	4	0.12881	0.27	7.03	2.4	0.396	2.4	0.994
13.1	Elongate in matrix	0.02	742	3575	4.98	1.00	244	2088	± 49	2069	± 3	-	0.12787	0.16	6.75	2.8	0.383	2.8	0.998
SB389 - N	Aount Z6389																		
6.1	Elongated in matrix Bt	0.01	3715	15301	4.25	1.2	1267	2156	± 41	2072	9∓	-5	0.128072	0.33	7.012	2.3	0.3971	2.2	0.989
6.2	Elongated in matrix Bt	0.01	3998	15206	3.93	1.4	1340	2124	± 51	2073	9∓	ų	0.128142	0.37	6.895	2.8	0.3902	2.8	0.992
7.1	In matrix	0.01	5536	22675	4.23	0.8	2049	2310	± 41	2075	± 2	-13	0.128300	0.12	7.623	2.1	0.4309	2.1	0.998
10.1	In matrix	0.01	5388	25521	4.89	1.2	1824	2142	± 41	2074	± 5	-4	0.128265	0.26	6.969	2.3	0.3941	2.3	0.993
11.1	Elongate in matrix	<0.01	5967	29638	5.13	0.4	2055	2173	±36	2072	± 3	-6	0.128131	0.16	7.083	1.9	0.4009	1.9	0.997
14.1	In matrix Ms	0.00	3748	15109	4.16	0.7	1290	2171	±45	2075	± 3	-5	0.128339	0.15	7.087	2.5	0.4005	2.4	0.998
Errors are	1-sigma; Pbc and Pb* indicate the commo	n and radioge	anic portions	, respectively	. (1) Com	mon Pb cor	rected using	measured 2	04Pb.										

Error in Standard calibration was 0.83 % (SB398 analyses) and 0.69% (SB389 analyses) (not included in above errors above but required when comparing different mounts)

for one of the monazites dated in SB398. (c) Results of SHRIMP analyses for SB398 on Concordia diagram with ellipses showing decay constant uncertainty. (d) Coarse-grained matrix monazite in sillimanite porphyroblast, elongated parallel with tectonic fabric in the matrix of SB389. (e) Results of SHRIMP analyses for sample SB389 on Concordia diagram with ellipses used to calculate age intercept. (f) Weighted mean average of U-Pb analyses for sample SB389.

are elongated parallel to the main S_2 fabric or sub-parallel to kyanite porphyroblasts (Fig. 4.13a; b; c). Eighteen dates were derived from monazite grains in SB398, yielding an upper intercept age of 2073 ± 2 Ma (MSWD = 0.72, P = 0.78, Fig. 4.11c) and a weighted mean 207 Pb/ 206 Pb age of 2073 ± 2 Ma (MSWD 0.68, P = 0.83) (Table 4.4). The ca. 2073 Ma intercept age is interpreted to record the timing of the growth of kyanite, with metamorphism occurring at the transition from amphibolite to granulite facies immediately above the solidus.

4.5.2.2 Sample SB389 (biotite-sillimanite gneiss)

Monazites in sample SB389 occur in the matrix commonly associated with biotite and aligned with the penetrative tectonic fabric (Fig. 4.13d). Back-scattered electron images show rare patchy compositional zoning and large fractures. Monazites in sample SB389 show two distinct morphologies. One population is subhedral to euhedral coarse grains (up to 200 μ m along the c-axis). These occur in either the matrix or biotite porphyroblasts. The other population are elongate, sub-rounded grains, often with embayed grain boundaries. They are often hosted within large fibrolitic sillimanite pseudomorphs (Fig. 4.13e, 4.13f). Six U-Th-Pb analyses were performed on five grains, representative of the two morphologies, yielding an upper intercept age of 2073 ± 6 Ma (MSWD = 0.18, P = 0.95) and a weighted mean ²⁰⁷Pb/²⁰⁶Pb age of 2074 ± 3 Ma (MSWD = 0.18, P = 0.97) (Table 4.4). Monazites in all textural locations yielded identical dates. As the monazites in SB389 were not shielded in porphyroblasts, this age likely represents the timing of the host rock passing through the closure temperature of monazite as the rock returned to subsolidus conditions during cooling and exhumation.

4.6 Discussion

4.6.1 Metamorphic evolution of the study area

The culmination of an extensive, regional-scale field mapping campaign, petrological studies and thermodynamic calculations of P-T conditions and P-T path determinations are presented in Figure 4.14, revealing distinct tectono-metamorphic domains and significant, lateral metamorphic breaks along the major shear zones bounding the Sefwi greenstone belt and normal-sense detachment faults. In this section, we summarise the progression of metamorphism within the study area, revealing

higher pressures and a more complex metamorphic history than previous studies suggest.

- 1. We interpret the earliest prograde metamorphic event from phases preserved as plurimineral aggregates (Chl + Pl + Rt + Qz \pm Bt) in garnet cores of the Grt-Ky-Rt paragneiss sample (SB398) from the Chiraa high-grade domain. This is interpreted as the partial preservation of an early prograde assemblage at temperatures below ~590 °C, in the stability field of rutile (P > 10 kbar). Whilst precise pressures and temperatures could not be constrained, this assemblage is associated with epidote-amphibolite facies conditions and a cool apparent geothermal gradient of <15 °C km ⁻¹. SB398 is the only sample to record such conditions in the early stage of the metamorphic evolution of the study area. This event is the earliest metamorphic event, however, it is relatively poorly constrained and is therefore denoted as M1e.
- 2. Subsequent prograde high-pressure amphibolite facies conditions (10.0–11.5 kbar, 600–650 °C) are deduced from phases in the garnet mantle of SB398 (Fig. 4.10b, 4.10d) in the Chiraa high-grade domain and from garnet porphyroblast compositions of the garnet amphibolite (SB344; Fig. 4.12b), ~150 km southwest, in the adjacent Kukuom-Juaboso medium-grade domain. From these two samples, we interpret metamorphism associated with an apparent geothermal gradient of 15–17 °Ckm⁻¹ that was contemporaneous across the two domains and part of the same metamorphic event (M1). The Opx-free, Grt + Cpx + Pl ± Hbl mafic granulite (SB177) on the southern margin of the Chiraa domain indicates higher, but poorly constrained, high pressure granulite facies metamorphic conditions (>10 kbar, >700 °C), associated with an apparent geothermal gradient of 11/2 transition.
- 3. In contrast to the samples above, sample SB389 in the Chiraa high-grade domain, ~20km east of SB398, indicates different initial conditions. Relict garnet porphyroblasts in biotite sillimanite paragneiss sample SB389, indicate initial conditions of 5–8 kbar, 560–640 °C. These conditions indicate a higher apparent geothermal gradient than those of M1, however, due to evidence of extensive diffusional reequilibration, we suggest that such constraints are not significantly robust for further interpretation.
- 4. Following M1, the next metamorphic event, M2, was characterized by metamorphic conditions of 8–10 kbar, 650–700 °C, establishing a new apparent geothermal gradient

of 20–30 °C km⁻¹. This conditions are associated with anatexis recorded only in samples from the Chiraa high-grade domain, best reflected in samples SB389 and SB398 (Fig. 4.10c, 4.11c). *P-T* conditions for these two samples converge during M2 metamorphism. Sample SB398 records well the *P-T* path, evolving from M1 to M2 through heating during minor decompression (<2 kbar) (Fig. 4.10). This compares with sample SB389 (Fig. 4.11) 20km east, that evolves from initial medium-pressure stages and converges towards these same conditions by heating and burial (Fig. 4.9). If these two paths are assumed to be contemporaneous it suggests that the entire terrane is heating up while parts of it are being exhumed (SB398) and others being buried (SB389).

- 5. In-situ monazite U-Pb SHRIMP ages of ca. 2073 Ma were yielded by both paragneiss samples (SB398 and SB389), hosted in kyanite porphyroblasts, matrix biotite and as individual grains within the matrix. The consistency of the U-Pb dates indicates relatively uniform monazite crystallisation within the Chiraa Domain. This suggest convergence of *P-T* paths discussed in point 4. This age post-dates zircon U-Pb crystallization ages of ca. 2092 Ma for two mica granites and pegmatites within the domain (Agyei Duodo *et al.*, 2009; Petersson, *et al.*, 2016), supporting the concept of zircon ages tending to be older than monazite ages in the same rock, commonly observed in high-grade metasediments (Kelsey, *et al.*, 2008). We propose that the uniform monazite ages records the monazite closure temperature at the beginning of the retrograde paths thus recording the initialtion of colling and exhumation.
- 6. Overprinting M2 metamorphism, M3 metamorphism is characterised by parallel retrograde *P-T* paths of samples located in the Chiraa high-grade domain (SB398 and SB389), which reach average amphibolite facies conditions of 5.5–7.0 kbar and 560–620 °C (SB398: Fig. 4.10c, d; SB389: Fig. 4.11c, d). This is interpreted as exhumation of the entire domain during M3. Neither *P-T* paths nor petrographic relationships in these samples indicate isothermal decompression, instead, they follow a moderately steep gradient with decreasing pressures and temperatures. Preservation of these assemblages suggests minimal fluid interaction in the later stages of the metamorphic history.



Fig. 4.14. a) Metamorphic map of the Sefwi Greenstone belt and adjacent metasedimentary provinces with calculated P-T paths. The migmatitic Chiraa domain is located in the northeast, separated from the amphibolite facies to locally migmatitic domains of the central portions of the Sefwi belt by the reactivated Kenyase-Yamfo fault. The greenschist facies volcano-plutonic Sefwi-Wiawso domain represents the southern portion of the belt, tectonically bounded by the reactivated Ketesso shear zone to the north and the Afema shear zone to the south. Fault kinematics indicate latest movement with the exception of the sinistral strike-slip component of D_2 inferred from structural and metamorphic relationships; b, c, d) Synthesis of metamorphic paths of high-metamorphic rocks exposed in SW Ghana with geothermal gradients and the aluminosilicate triple junction as a point of reference, revealing an initial apparent geothermal gradient of 15 - 17 °C km⁻¹ recorded by samples in multiple tectono-metamorphic domains. Subsequent partial melting is associated with sustained high pressures as the rocks crossed over the solidus. Note the distinct difference between the geothermal gradients recorded by retrograde paths between the anatectic samples (SB389; SB398) and the garnet amphibolite (SB344).

- 7. Southwest of the Chiraa domain in the Kukuom-Juaboso domain, the garnet amphibolite sample SB344 of point 2 above shows a different metamorphic history after the high-pressure metamorphism of M1. It does not record an increase in temperatures after peak temperatures. It instead records cooling during exhumation, associated with an overprinting high-pressure greenschist facies metamorphism (6.5–9 kbar, 450–500°C) (Fig. 4.12c). This is interpreted as a divergence in the metamorphic histories of the two terranes, occurring at the M2/M3 transition.
- 8. The latest recognisable metamorphic overprint is interpreted as M4. Continued retrogression to greenschist facies conditions is localized along tectonically-reactivated high-strain zones and domain boundaries (Fig. 4.8f; Fig. 4.14). On the southern margin of the Chiraa domain, the peak assemblage of mafic granulite sample SB177 is overprinted by an assemblage typical of lower greenschist facies (Chl + Ab + Cal + Ep + Qz).

Figure 4.14 presents a synthesis of the *P*-*T* paths and *P*-*T* conditions recorded by high-grade rocks in SW Ghana. In summary, bringing together all the available metamorphic data derived from these samples, we suggest that the history of metamorphism for the belt is: an initial crustal thickening event (D_1), responsible for the burial of supracrustal rocks and the M1e and M1 assemblages; subsequent heating during D_2 is only recorded in the Chiraa high-grade domain with western samples indicating heating during exhumation and eastern samples indicating heating during burial, demonstrated by the M2 phase assemblages; exhumation during cooling occurs in both the Chiraa and the Kukuom-Juaboso domain during D_3 , however, this occurs under contrasting apparent geothermal gradients between the domains suggesting heterogeneous metamorphic histories exist within the study area. The tectonic juxtaposition of the Chiraa and Kukuom-Juaboso domain with the adjacent, greenschist-facies, chlorite-white mica schists of the Sunyani-Comoé domain to the west is interpreted as the product of exhumation during D_3 .

The determined metamorphic history is significantly more complex than that deduced by Galipp, *et al.* (2003), and indicates much higher pressures than metamorphic conditions (490–580 °C, 4–6 kbar) of Galipp, *et al.* (2003) derived from garnet-biotite thermometry and plagioclase-amphibole thermobarometry for a schist and a mafic meta-granitoid, located ~ 35km north-northeast of SB344 in the Kukuom-Juaboso domain. Galipp, *et al.* (2003) infer a clockwise *P-T* path from an overprinting retrograde assemblage of Chl + Ep + Ab + carbonate, similar to petrological observations the retrograde assemblage in mafic granulite sample SB177.

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4.6.2 Regional context

Geochronological constraints of metamorphism and magmatism in the Sefwi belt are severely limited, hindering correlation with similar greenstone or high-grade terranes of the West African Craton. Emplacement of hornblende-biotite bearing granitoids in domains of the Sefwi belt is restricted to \sim 2180 – 2150 Ma (Delor, *et al.*, 1992; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2003; Hirdes, *et al.*, 2007), with late high-K granites and granodiorites emplaced at ca. 2136 Ma (Amponsah, 2012; Hirdes, *et al.*, 2007). These ages are significantly older than metamorphic ages revealed in this study and, therefore, are unlikely to be correlated. A single metamorphic date exists in the Ivorian extent of the Sefwi belt, with a metamorphic titanite in a dioritic amphibolite yielding a Pb-Pb age of 2144 ± 3 Ma (Hirdes, *et al.*, 2007). Again, this age is significantly older than monazite ages detailed in this study, however, in the absence of a structural context, no further correlation can be made.

U-Pb zircon crystallisation ages of 2092 ± 2 Ma (Agyei Duodu, *et al.*, 2009), 2093 ± 2 Ma and 2092 ± 4 Ma (Petersson, *et al.*, 2016) were yielded by exposures of two-mica granite and pegmatite within the Chiraa high-grade domain, obtained using SIMS analysis. Leucogranites in the adjacent Sunyani-Comoé domain were emplaced parallel to the northwest margin of the Sefwi belt in upper crustal, greenschist facies metasedimentary rocks between 2088 ± 1 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992) and 2081 ± 6 (Hirdes, *et al.*, 2007) (locations in Fig. 4.12), along strike of the Chiraa domain. The timing of peak to post-peak metamorphism in the Chiraa high-grade domain, as constrained by U-Pb monazite ages, suggests temperatures in the Chiraa domain remained above monazite closure temperatures until ca. 2073 Ma. As the Chiraa domain and the Sunyani-Comoé domain are now exposed at the same crustal level, it is reasoned that tectonic juxtaposition by differential exhumation post-dates both monazite growth in the Chiraa domain and the along-strike emplacement of leucogranites in low-grade metasediments of the Sunyani-Comoé domain.

In the context of southern Ghana, MP-MT metamorphic conditions and peak temperatures of ~550 °C in the neighbouring Ashanti greenstone belt occurred at ca. 2100 Ma, with post-peak amphibolite facies metamorphism constrained by a hydrothermal xenotime crystallisation SHRIMP U-Pb age of 2063 \pm 9 Ma (Pigois, *et al.*, 2003). Staurolite-grade, amphibolite facies metamorphism of the Tarkwa Phyllite in the Damang gold deposit, in the southern central region of the Ashanti belt, was constrained by EMPA U-Pb monazite analysis, which yielded an age of 2005 Ma (White, *et al.*, 2014), however, the structural context of the monazite may be obscured by textures related to late dyke intrusion. Further south, migmatitic orthogneisses are documented in the Kibi-Winneba belt and in the far southeast of the Ashanti belt (Fig. 4.2) with calculated peak metamorphic conditions of 5 – 8.5 kbar, ~600 °C, which are as yet undated (Feybesse, *et al.*, 2006; Opare-Addo, *et al.*, 1993).

Exposed elsewhere in the West African Craton, high-grade terranes comparable to the Chiraa domain reveal metamorphic ages significantly older than ages in southern Ghana. Garnet-whole rock Sm-Nd isochron ages constrain the timing of the amphibolite facies, Yaouré ortho- and paragneisses in central Ivory Coast at 2153 \pm 13 Ma, directly west of the study area (Boher, *et al.*, 1992). North of the study area, paragneisses and chlorite-white mica schists in NW Ghana and eastern Burkina Faso, respectively, reveal relic blueschist facies mineral assemblages, matching low apparent geothermal gradients (10 – 15 °Ckm⁻¹) (Block, et al., 2015; Ganne, et al., 2012). The timing of subsequent anatexis at the amphibolite-granulite transition and a late amphibolite facies overprint in the high grade rocks of NW Ghana occurred between ca. 2140 and 2110 Ma, as indicated by zircons and monazite ages yielded by magmatic and metamorphic units (Block, et al., 2015; 2016b). Block, et al. (2015) elucidate a metamorphic history analogous to that of this study, however, the discrepancy in the timing of metamorphism suggests high-grade metamorphism occurred ~40-70 m.y. earlier in NW Ghana than in SW Ghana. High-pressure granulite facies assemblages in western Ivory Coast post-date Ghanaian high-grade metamorphism, forming at c. 2030 during the terminal collision between the Archaean and Palaeoproterozoic domains of the WAC along the Sassandra Fault (Fig. 4.2) (Grt-WR Sm-Nd age, Kouamelan, et al., 1997; Pitra, et al., 2010). Whilst further geochronological data is required to fully understand craton-wide metamorphic relationships, the considerable discrepancy in the timing of high-grade metamorphism and lower crustal exhumation between NW Ghana, SW Ghana and western Ivory Coast suggests a diachronous metamorphic history for the Eburnean Orogeny during the assembly of the West African Craton.

4.6.3 Burial and exhumation

This study reveals an initial period of crustal thickening and burial of supracrustal rocks in high-grade domains to depths greater than 35 km (10–11 kbar) (Fig. 4.14). M1 is characterised by intermediate high-*P/T* conditions, matching a cool apparent geothermal gradients of 15–17 °km⁻¹, suggesting rapid burial of sediments during crustal thickening. It has been well established that similar low apparent geothermal gradients (10–20 °Ckm⁻¹) record the thermal conditions associated with burial of crustal material to significant depths during the early stages of continental collision (England & Richardson, 1977; England & Thompson, 1984). Given the paucity of outcrop and the extensive overprinting by

subsequent transcurrent tectonism, evidence for the crustal thickening mechanism of D_1 is largely obscured. However, based on the ubiquitous bedding-parallel metamorphic foliations and rare, moderately to steeply, north and south pitching down-dip mineral stretching lineations on E-W to ENE-WSW–striking early shear zones, D_1 is likely characterised by N-S to NNW-SSE shortening and thrust-related thickening in a convergent tectonic setting. Similar crustal thickening and rapid burial of metasediments is recorded in northwest Ghana (Block, *et al.*, 2015).

Investigation of negative dP/dT segments of *P*-*T*-*t*-*D* paths aids in the understanding of the evolution of orogens. They lend additional insight into the uplift or exhumation mechanism of metamorphic terranes, including exhumation of high-pressure and ultra-high-pressure (HP and UHP) rocks via erosion, extrusion, corner flow, buoyancy and extension (Franěk, *et al.*, 2011; Guillot, *et al.*, 2009; Platt, 1993; Teyssier & Whitney, 2002), each recording a unique *P*-*T*-*t* path. Combined metamorphic and structural data presented in this study indicate a progressive transition from D₁ to D₂ under continued NNW-SSE shortening. Following peak-*P* conditions, sample SB398 of the Chiraa domain underwent heating and small amount of exhumation (<2 kbar) to suprasolidus conditions, with the negative dP/dT segment of the clockwise *P*-*T* path likely recording thermal relaxation of the crust (England & Thompson, 1984) and the termination of crustal thickening.

Stromatic leucosomes, parallel with the primary gneissosity or bedding-parallel S_1 foliations, feed into NNE-striking dilational zones, indicating late D_1 —syn- D_2 partial melting. D_2 structures are characterised by ~E-W to NE-SW striking, upright to slightly inclined, tight F_2 folds, seen refolding S_0/S_1 at an outcrop scale and D1 shear zones a regional scale (Fig. 4.15). The L_2 mineral stretching lineation is variably defined by kyanite and amphibole/biotite in the Chiraa and Kukuom-Juaboso domains, respectively. Both F_2 fold axes and L_2 mineral stretching lineations predominantly pitch shallowly W to WSW, or alternatively, shallowly to the east due to later deformation. The dip direction of L_2 and F_2 is oblique to the NE-SW orientation of the major shear zones bounding the Sefwi Belt. The coaxial relationship of F_2 and L_2 stretching lineations, indicating constrictional folding, and their obliquity to the regional shear zones are characteristic of transtension tectonic regimes (Fossen, *et al.*, 2013). The orientation of the major tensional axis of the strain ellipsoid during D_2 , is therefore interpreted as ENE-WSW, marking the transition from a NNW-SSE compressive regime to a ENE-WSW tensile regime.

Metamorphic breaks coincide with the Kenyase-Yamfo and the Ketesso high-strain zones, with a regional decrease in metamorphic grade from the northeast to the southwest. The most significant metamorphic breaks, however, occur along NNE-striking detachment faults. We propose that anatectic lower crustal rocks of the Chiraa high-grade domain and amphibolite facies, middle crust of the Kukuom-Juaboso domain were differentially exhumed under sinistral transtension and tectonically juxtaposed against the greenschist facies metasediments of the Sunyani-Comoé domain. The development of a kilometric drag fold rotating the tectonic fabrics in the Chiraa high-grade domain is consistent with sinistral displacement along the high strain zone marking the southern boundary of the domain (Fig. 4.15).



Fig. 4.15. a) Metamorphic map with foliations trajectories of the Chiraa high-grade domain showing rotation of the S1 and S2 foliations within the domain, and subsequent overprinting and transposition

of foliations along major dextral high-strain zones during D_3 E-W shortening; (b) Kyanite-defined L2 stretching lineation plunging ~30° towards N 273°; (c) P-T paths and metamorphic conditions highlight D2 exhumation.

Syn-orogenic partial melting of the lower crust has been recognised as the catalyst for the rheological weakening of the crust, resulting in crustal flow in response to either tectonic or gravitational stresses (Brown, 2010; Royden, *et al.*, 2008; Vanderhaeghe, 2009; Vanderhaeghe & Teyssier, 2001).

Evidence of partial melting in the lower crustal rocks of the Chiraa domain during M_2/D_2 , likely contributed to rheological weakening of the over-thickened crust, facilitating the transition to the D₂ transtensional regime orthogonal to the shortening regime of D₁, after which exhumation occurred. Concurrent cooling and decreasing pressure after peak temperatures is evident in steep negative dP/dT segments of P-T paths for samples from the Chiraa domain. The gradient of their retrograde paths crosses steady state geotherms suggesting exhumation exceeded thermal reequilibration, thus precluding homogeneous uplift and erosion of the terrane. In Phanerozoic orogenic belts, gravitationally-driven orogenic collapse is associated with anatectic dome structures, downdip mineral stretching lineations and isothermal decompression in suprasolidus segments of P-T paths (Teyssier & Whitney, 2002; Whitney, et al., 2004), similar to those documented in high-grade domains of Palaeoproterozoic NW Ghana (Block, et al., 2015; Block, et al., 2016b). The shallow to moderate pitches of L₂ stretching lineations and the association of small temperature decreases with decreasing pressure during exhumation observed in SW Ghana distinguishes transtension-related exhumation from coupled extension and gravitational collapse driving exhumation in NW Ghana. We note that gravitational forces acting on the over-thickened crust likely assisted for the transition from a compressive to tensile regime, resulting in orogen parallel flow and late orogenic exhumation reminiscent of modern orogenic processes, however, we are reticent to invoke gravitational collapse based on the discussed metamorphic constraints.

Transpressional dextral reactivation and transposition of pre-existing tectonic contacts to subvertical shear zones along domain boundaries are attributed D_3 E-W bulk shortening. Pure shearrelated structures forming under E-W shortening are evident in anisotropic units striking obliquely to the major NE-oriented crustal architecture. A penetrative, N-S striking, crenulation cleavage and upright, open F₃ folds overprint S₁ and S₂ decreasing in intensity towards the south. Retrograde greenschist facies metamorphism is localised along shear zones, defined by chlorite-white mica S-C or S-C' fabrics.

4.6.4 Implications for Palaeoproterozoic tectonics

The Palaeoproterozoic crust exposed in SW Ghana and SE Ivory Coast comprises small high grade domains, metamorphosed at high-pressure amphibolite to transition amphibolite-granulite facies metamorphic conditions. These domains are tectonically juxtaposed against low-grade, upper crustal metasediments along detachments faults and regional scale shear zones. The presence of strong metamorphic breaks does not conform to the spatially uniform metamorphic conditions and homogeneous uplift proposed by some authors in Precambrian accretionary or hot orogens (Cagnard, *et al.*, 2006; Gapais, *et al.*, 2009), attributed to warmer, weak lithosphere (Cagnard, *et al.*, 2011). The metamorphic record of the study area instead indicates early heterogeneous crustal thickening, associated with cool apparent geothermal gradients, reminiscent of modern incipient collisional orogenesis.

Numerical modelling of post-accretionary deformation in wide, hot orogens displays lateral, orogen parallel flow is produced during bulk constriction, common to Precambrian hot orogens and modern orogenic plateaux (Cagnard, *et al.*, 2006; Chardon, *et al.*, 2009; Chardon, *et al.*, 2011; Cruden, *et al.*, 2006). Here, however, we illustrate the development of constrictional structures occurred under moderate geothermal gradients of 20–30 °Ckm⁻¹ resulting in differential exhumation of anatectic mid-to lower crustal blocks, reminiscent of modern processes. Such evidence suggests that transtension was instead driven by combined boundary forces, likely associated with a period of oblique plate divergence, with the assistance of gravitational forces acting on the thickened, partially molten orogenic crust. Similar crustal mechanical behaviour in the Palaeoproterozoic lower crustal exhumation in SW Ghana is analogous to the Devonian transtension-related exhumation of ultra-high-pressure rocks of the Western Gneiss Region of the Scandinavian Caledonides (Fossen, *et al.*, 2013; Krabbendam & Dewey, 1998). These orogenic processes significantly contrast those observed in other Palaeoproterozoic accretionary orogens such as those of the southern Finnish Svecofennides (Cagnard, *et al.*, 2007).

Eclogites and high-P metamorphism have been documented in a number of Palaeoproterozoic orogens, including the Usagaren Orogen, Tanzania, at ca. 2000 Ma (Collins, *et al.*, 2004) and, more recently, in the Trans-Hudson Orogen at ca. 1800 Ma (Weller & St-Onge, 2017). These authors argue that this indicates modern subduction-related orogenic processes were active early in the Earth's history. The findings of this study do not reveal direct evidence of subduction-related processes. Instead, we propose that the combined evidence of HP-MT assemblages, low initial apparent geothermal gradients,

heterogeneous crustal thickening and the generation and emplacement of leucogranitic plutons parallel to the northwest margin of the Sefwi belt is the ancient expression of a Palaeoproterozoic collisional orogen suturing two micro-continental fragments represented by southern Ghana and central Ghana/ central Ivory Coast.

Our findings, in addition to other recent studies (e.g. Block, et al., 2015; Ganne, et al., 2012), document diverse geothermal gradients within the West African Craton, indicating a more heterogeneous tectono-metamorphic evolution than previously assumed. These results contrast with the high geotherms recorded during collisional orogenesis involving the juvenile crust of Palaeoproterozoic Southern Svecofennian Arc Complex, southern Finland (Cagnard, et al., 2007). It must be noted, however, that the in sporadic preservation of HP-LT or HP-MT metamorphic conditions within the WAC, pressures never exceed 14kbar, potentially indicating the maximum amount of crustal thickening supportable by the juvenile Palaeoproterozoic crust. Diverse geothermal gradients are associated with terrane accretion in the Archaean Barberton granitoid-greenstone terrain (Moyen, et al., 2006) and in throughout the eastern Yilgarn Craton (Goscombe, et al., 2009). Whilst granulite facies metamorphism is ubiquitous throughout Earth's history, low geothermal gradients (10-20 °Ckm⁻¹) associated with eclogite-high pressure granulite (E-HPG) metamorphism first appeared in the Neoarchaean, with an increasing number of occurrences in the Palaeoproterozoic and into the Phanerozoic (Brown, 2007, and references therein). Coupled with the occurrence of ultra-high temperature (UHT) metamorphism after the Archaean-Proterozoic boundary, Brown (2007; 2009) proposes that changes to global geodynamic processes created a proto-tectonic regime during the Neoarchaean, associated with an increased diversity in tectono-metamorphic environments. In this context, the assorted thermal regimes and tectonic processes recorded in the West African Craton, in the absence of extreme pressure metamorphism attributed to modern subduction-related orogenic processes, supports the notion of a transitional Proterozoic plate tectonic regime. Furthermore, this research is complementary to the growing body of research that suggest Palaeoproterozoic orogens forming between 2100 and 1800 Ma document growing heterogeneity in thermal regimes, whilst also displaying spatial and temporal variations in metamorphism and strain accommodation between individual Palaeoproterozoic provinces.

4.7 Conclusions

The metamorphic record of juxtaposed domains in the Palaeoproterozoic Sefwi greenstone belt of the

West African Craton reveals a far more complex metamorphic history than previously recognised. We document low apparent geothermal gradients and clockwise P-T-t paths with heterogeneous prograde segments. Sharp lateral metamorphic breaks are interpreted as the product of tectonic exhumation of crustal slices from different depths. Exhumation is associated with strain localisation along craton-scale ductile shear zones and normal sense detachments faults during oblique divergence, likely assisted by partial melting in the lower crust. Palaeoproterozoic rocks in southwest Ghana provide excellent insight into the link between high grade metamorphism within greenstone belts and the craton-scale shear zones that dominate the West African Craton. These results contrast the weak, hot orogen model proposed for some areas of the craton. The metamorphic record of the study area is interpreted as the product of monocyclic orogenesis. The discrepancies in the timing of high grade metamorphism throughout the craton, however, suggests episodic collisional orogenesis during the Eburnean Orogeny was ultimately responsible for craton assembly, superimposed on the accretionary orogenic system of the West African Craton.

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Appendix C. Pseudosection data files

Perplex data files for pseudosection calculation.

Refer to electronic data files with prefix "Ch4_AppendixC_Perplex"

Chapter V

Crustal evolution of the Sefwi Greenstone Belt: Implications for the West African Craton



5.0 Introduction

The previous research chapters explored the tectonic style and exhumation mechanism of the Eburnean Orogeny as preserved in SW Ghana. This chapter focuses on the magmatic evolution of the Sefwi Greenstone Belt in order to understand the geodynamic setting of crustal growth. It presents new petrological and geochemical analysis of representative magmatic suites from throughout the Sefwi Greenstone Belt and the Sunyani-Comoé Domain. Major and trace element data analysis investigates potential magma sources and petrogenetic processes responsible for the Palaeoproterozoic crust formation. Zircon U-Pb and Lu-Hf analyses constrain the timing of magmatism, as well as providing us with information on the timing of mantle extraction, crustal residence times, and the nature of the proto-crust. The timing of magmatism and $\varepsilon Hf_{(i)}$ values are compared to other areas in the West African Craton in order to investigate potential isotopic boundaries within the craton, providing crucial information on the craton architecture. Finally, we consider the broader geodynamic setting responsible for the formation and preservation of the West African Craton relative to Archaean and Phanerozoic geodynamic models.

Chapter 5: Palaeoproterozoic juvenile crust formation of SW Ghana in the West African Craton: New insights from igneous geochemistry and U-Pb-Hf zircon data

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Abstract

New geochemical and isotopic data are presented for the magmatic suites exposed in southwest Ghana in the south-eastern portion of the Palaeoproterozoic (2300–2070 Ma) West African Craton. This study focuses on the magmatic evolution of the Sefwi Greenstone Belt, re-defining the geochemical and isotopic characteristics of the major magmatic suites emplaced between ca. 2155 and 2136 Ma. The juvenile Palaeoproterozoic crust was derived from crustal material extracted from the depleted mantle between ca. 2650 and 2260 Ma, indicating short crustal residence times and includes minimal reworking of Archaean continental crust.

New whole-rock geochemical data reveals the wide range of petrological, geochemical and isotopic signatures of magmatic suites in southern Ghana, which indicate contemporaneous emplacement of magmas from a range of sources within a volcanic island arc setting. The five distinct magmatic suites include: 1) biotite-dominant and hornblende-dominant, SiO₂- and Na₂O-rich tonalites-trondhjemitegranodiorite (TTG) granitoids and associated dacites; 2) calc-alkaline, LILE-enriched diorites and associated andesitic lavas; 3) pyroxenite, gabbro and layered mafic intrusives; 4) high-K quartz monzonites; and, 5) incompatible element rich, two-mica granites and muscovite leucogranites. Trace element patterns indicate near ubiquitous negative Nb-Ta, P and Ti anomalies and positive Pb and Ba anomalies in the felsic suites, indicative of a crustal influence during magma genesis in an island arc setting. Diorites and biotite-bearing low-HREE TTGs are emplaced contemporaneously at ca. 2155 Ma, revealing contrasting sources. Diorite magmas are produced thorough the partial melting of the incompatible element-enriched metasomatised mantle wedge, whilst the biotite-bearing TTGs were likely produced through the partial melting of young, hot subducting oceanic crust at depth. The hornblende TTGs are attributed to the shallow, partial melting of mafic crust based on low Sr and high HREE concentrations, weak REE fractionation and major element ratios compared to experimental melts. The emplacement of high-K quartz monzonites at ca. 2136 Ma reveals mixing of mantlederived magmas and partial melts derived from existing TTGs in the crustal pile. The subsequent emplacement of two-mica granites and leucogranites between ca. 2092-2081 Ma is interpreted to be the product of crustal anatexis during an arc-arc collisional event, with a proposed suture parallel to the NW margin of the Sefwi Belt. In-situ zircon Lu-Hf analysis reveals consistently positive $\epsilon H f_{(t)}$ values between +4.9 and +8.0 for TTGs, +2.2 and +5.5 for diorites and +2.7 to +5.8 for dacitic lavas, highlighting a distinctly radiogenic source during crustal growth, potentially representing a juvenile

isotopic boundary within the craton. Subsequent emplacement of quartz monzonites reveals a wider range of $\varepsilon Hf_{(i)}$ values between +1.2 and +6.5, suggesting growing heterogeneity of magma sources. We hypothesise that the formation and accretion of the West African Craton reflects the rapid accumulation and amalgamation of individual island arc terranes generated by two coeval intra-oceanic subduction zones, representing an ancient expression of the Wilson cycle.

Key words: Palaeoproterozoic; subduction-collision; juvenile magmatism; Zircon Lu-Hf; West African Craton

5.1 Introduction

The Archaean and Proterozoic rock record reveals temporal changes in the mechanisms and processes associated with crustal growth and preservation. These changes reflect the secular cooling of the Earth, the differentiation of the Earth's crust and evolving plate dynamics (e.g. Davies, 2001; Hawkesworth & Kemp, 2006a; Herzberg, et al., 2010). Crustal growth models indicate that more than 50% of the cumulative volume of continental crust had formed by 3000 Ma (Belousova, et al., 2010; Cawood, et al., 2013; Dhuime, et al., 2012; Hirdes, et al. 1992; Taylor & McLennan, 1985), with continuous additions hypothesised from the early Earth through to the present day. However, major changes are recorded in the continental rock record between 3000 and 2500 Ma, including the geochemical evolution of volcanic and plutonic rocks (Keller & Schoene, 2012; Laurent, et al., 2014; Moyen & Martin, 2012), the diversification and secular evolution of the metamorphic record (Brown, 2006; Brown, 2007a) and changes in orogenic strain patterns, crustal architecture and rheology (Cagnard, et al., 2011; Chardon, et al., 2009; Rey & Houseman, 2006). This multi-disciplinary body of evidence signals significant changes in global geodynamics, often interpreted as the commencement of some form of plate tectonics (Condie & Aster, 2010; Condie & O'Neill, 2010; Keller & Schoene, 2012; Sizova, et al., 2010; Windley, 1984). Indeed the onset of plate tectonics is likely responsible for the transition from net crustal growth to major lithospheric recycling and reworking as part of the Wilson cycle, with the destruction or consumption of tectonic plates in subduction zones (e.g. Cawood, et al., 2013; Dhuime, et al., 2012).

A number of datasets within the geological record illustrate distinct clusters of magmatic and tectonic activity during the evolution of the early Earth. Episodic crustal growth is suggested by peaks in the juvenile crust record at 2700, 1900 and 1200 Ma (Cawood, *et al.*, 2009; Condie, 1995; Condie, 1998); the orogenic record displays discrete events at ca. 3800–3500, 2900–2600, 1900–1600, and 1200–900 Ma (e.g. Zhao, *et al.*, 2002), whilst the U-Pb detrital zircon ages define peak clusters at 2700, 2500, 2010, 1840, 1650, 1150 and 600 Ma (Condie, *et al.*, 2005; Condie, *et al.*, 2011). These peaks often coincide with supercontinent formation (e.g. Condie & Aster, 2010; Nance, *et al.*, 1988; Rogers & Santosh, 2004), however, they all represent periods during which conditions were favourable for preservation of the crust (Hawkesworth, *et al.*, 2009; 2010). Falling between these maxima, cratonic domains of the West African-Sao Luis Craton and Guiana Shield record a period of significant juvenile crustal growth during the early Palaeoproterozoic (2300–2070 Ma), representing one of the youngest cratonic provinces on Earth (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992; Gasquet, *et al.*, 2003; Gruau, *et al.*, 1985; Klein, *et al.*, 2005; Taylor, *et al.*, 1992). Production and preservation of juvenile crust is attributed to the Eburnean–Trans-Amazonian Orogen (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992; Bonhomme, 1962), forming one of the earliest orogenic events of the 2100–1800 Ma period of global orogenesis that heralded the rise of the Palaeoproterozoic supercontinent Nuna (also called Columbia (Betts, *et al.*, 2016; Rogers & Santosh, 2002; Zhao, *et al.*, 2002). Juvenile Palaeoproterozoic rocks of the southern portion of the West African Craton, forming the Baoulé-Mossi Domain, comprise narrow, elongate- to arcuate greenstone belts alternating with parallel sedimentary basins or vast granite-gneiss domains of tonalite-trondhjemite-granodiorite plutons, bearing striking lithological and architectural similarities with Archaean granite-greenstone provinces, including the Superior Province of the Canadian Shield (Card, 1990a; Condie & Benn, 2006; Daigneault, *et al.*, 2003; Zeh, *et al.*, 2009; Zeh, *et al.*, 2013), and the Kola and Karelian provinces of the Baltic Shield (Sorjonen-Ward & Luukkonen, 2005, and references therein). Parallel greenstone belts, striking between N- and NE, occur across the craton and are often fault-bounded by regional to craton-scale shear zones.

The use of *in-situ* isotopic analysis of zircon, including both U-Pb and Lu-Hf systematics yield both the crystallisation age and the timing of extraction of the source material from the mantle (Hawkesworth & Kemp, 2006b; Patchett, *et al.*, 2004). This is a well-established method used to understand crustal evolution, crustal architecture and the geodynamic setting of crustal formation (e.g. Belousova, *et al.*, 2010; Cawood, *et al.*, 2013; Condie, *et al.*, 2005; Dhuime, *et al.*, 2012; Griffin, *et al.*, 2004; Griffin, *et al.*, 2000; Hawkesworth & Kemp, 2006b; Kemp, *et al.*, 2007).

This study reports new whole-rock geochemical data for twenty-nine igneous samples and ion microprobe (SHRIMP II) U-Pb and LA-MC-ICP-MS Lu-Hf zircon data for four samples. The selected samples are believed to be representative of the major magmatic suites exposed in the Sefwi Greenstone Belt and Sunyani-Comoé of southwest Ghana. The aim of this study is to examine the magmatic suites emplaced in southern Ghana and elucidate the crustal evolution of an archetypal Palaeoproterozoic greenstone belt. Furthermore, we discuss the implications of our results, combined with published data, within the context of the crustal architecture of the West African Craton and the geodynamic setting responsible for the formation and preservation of the juvenile crustal domain.

5.2 Palaeoproterozoic geology of West Africa

The West African Craton (WAC) encompasses the northern Reguibat Rise and the sub-Saharan Leo-Man Rise, each comprising 3600–2600 Ma Archaean provinces to the west and 2300–2070 Ma Palaeoproterozoic domains to the east (Abouchami, *et al.*, 1990; Potrel, *et al.*, 1998; Rollinson, 2016; Thiéblemont, *et al.*, 2001). The domains are separated by the overlying Mesoproterozoic-Palaeozoic Taoudeni Basin. The Archaean Man Shield is juxtaposed with Palaeoproterozoic rocks of the Baoulé-Mossi Domain along the Sassandra Shear Zone (Fig. 5.1) (Attoh & Ekwueme, 1997; Bessoles, 1977), with the Palaeoproterozoic Kayes and Kédougou-Kéniéba Inliers lying to the northwest. The Baoulé-Mossi domain is dominated by rocks of the Birimian Supergroup (2300 – 2125 Ma), forming linear to arcuate, N- to NE-trending volcanic greenstone belts and broad volcano-sedimentary domains, often bounded by craton-scale shear zones and separated by vast granite-gneiss terranes (Hirdes, *et al.*, 1996; Junner, 1940; Leube, *et al.*, 1990). Volcanic belts are commonly unconformably overlain by fluvio-deltaic sequences of the Tarkwa Group and equivalents (2107–2097 Ma) (e.g. Baratoux, *et al.*, 2011; Perrouty, *et al.*, 2012; Pigois, *et al.*, 2003; Zitzmann, *et al.*, 1997). The Palaeoproterozoic rocks were accreted and cratonised during the 2150–2070 Ma Eburnean Orogeny (Bonhomme, 1962; Feybesse, *et al.*, 2006; Oberthür, *et al.*, 1998).

Volcanic rocks of the Birimian Supergroup range in composition from basalt to rhyolite, displaying tholeiitic to calc-alkaline affinities (e.g. Baratoux, *et al.*, 2011; Hirdes, *et al.*, 1993; Senyah, *et al.*, 2016; Sylvester & Attoh, 1992), with subordinate intercalations of volcaniclastic or immature sedimentary rocks. Volcano-sedimentary domains comprise deformed sequences of greywacke, shale, argillitic and volcaniclastic rocks (pyroclastic and epiclastic), and rare chemical sedimentary rocks and calc-alkaline volcanic sequences (Hirdes, *et al.*, 1993; Leube, *et al.*, 1990). The volcanic and volcano-sedimentary rocks are extensively intruded by voluminous plutonic rocks forming a wide range of magmatic suites, with diorite to tonalite-trondhjemite-granodiorite (TTG), high-K granite, and leucogranite compositions. Prior to the advent of precision geochronological methods, plutonic rocks in Ghana were classified as "belt-type" or "basin-type", based on their host terranes. In southern Ghana, based on the predominant ferromagnesian mineral and alkali index classification, further subdivision of felsic plutons resulted in the four following suites: Dixcove, Winneba, Cape Coast and Bongo (Leube, *et al.*, 1990). The Dixcove granitoids are variably foliated, metaluminous, hornblendebearing plutons which intrude volcanic belts, whilst the biotite- and/or muscovite-bearing Winneba and Cape Coast granites and granodiorites are restricted to volcano-sedimentary domains or basins (Leube, *et al.*, 1990). Finally, the K-rich Bongo granites intrude Tarkwaian-like sediments in NW Ghana (Abitty, *et al.*, 2016; Leube, *et al.*, 1990).



Fig. 5.1. Simplified map of the West African Craton showing the Archaean Kénéma-Man domain and the Palaeoproterozoic Baoulé-Mossi domain, comprising volcanic greenstone belts (dark green), meta-sedimentary basins (light green) and large granitoid domains (orange). Yellow areas represent late fluvio-deltaic sequences of the Tarkwa Group (modified after BRGM SIGAfrique map, Milési, et al., 2004).

Mafic to intermediate volcanism occurred across the craton between 2300 and 2180 Ma (e.g. Agyei Duodu, *et al.*, 2009; Delor, *et al.*, 1995; Feybesse, *et al.*, 2006; Hirdes & Davis, 1998; Lompo, 2009), approximately coeval with the earliest pulse of TTG magmatism (de Kock, *et al.*, 2011; Gueye, *et al.*, 2007; Hirdes, *et al.*, 1992; Oberthür, *et al.*, 1998; Siegfried, *et al.*, 2009). The most significant volumes of syn- to late-tectonic granitoid intrusions, however, were emplaced near-continuously between 2150 and 2100 Ma, yet preserved no clear lateral age gradients (Agyei Duodu, *et al.*, 2009; Doumbia, *et al.*, 1998; Egal, *et al.*, 2002; Gasquet, *et al.*, 2003; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 1996; Oberthür, *et al.*, 1998; Tapsoba, *et al.*, 2013; Tshibubudze, *et al.*, 2013). Granitoids and lavas emplaced between 2100 and 2060 Ma are largely restricted to the Kédougou-Kéniéba Inlier and along the Archaean-Palaeoproterozoic contact (Egal, *et al.*, 2002; Gueye, *et al.*, 2007; Hirdes & Davis, 2002a),

with a significant, but localised population documented in southern Ghana along the northwest margin of the Sefwi Belt (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007).

Many authors liken the evolution from bimodal tholeiitic volcanism to calc-alkaline volcanism with the development of a modern-day oceanic island arc setting (Ama Salah, et al., 1996; Arndt, et al., 1997; Baratoux, et al., 2011; Béziat, et al., 2000; Dampare, et al., 2008; Soumaila, et al., 2008; Sylvester & Attoh, 1992). Other authors attribute the wide-spread tholeiitic basaltic volcanism and associated plutonic activity to the plume-related formation of an oceanic plateau (Abouchami, et al., 1990; Augustin & Gaboury, 2017; Boher, et al., 1992; Lompo, 2009). Alternatively, authors have invoked intra-cratonic rifting leading to the opening of oceans and the formation of accretionary systems (Leube, *et al.*, 1990) or a transtensional back-arc basin setting (Vidal & Alric, 1994). Isotopic studies of the juvenile terranes of the Baoulé-Mossi domain that formed between 2200 and 2100 Ma reveal Nd-model ages within ~300 Myrs of their magmatic crystallisation ages, recording crustal residence ages between 2500 and 2200 Ma (Abouchami, et al., 1990; Boher, et al., 1992; Gasquet, et al., 2003; Pawlig, et al., 2006; Peucat, et al., 2005). Nd and Hf-model ages reported in a small number of studies from far south-eastern Ghana and in the western portion of the Baoulé-Mossi domain, however, suggest a greater Archaean crustal influence or reworking of older Archaean crust in some portions of the domain (Parra-Avila, et al., 2016; Petersson, et al., 2017; Petersson, et al., 2016; Taylor, *et al.*, 1992).

The crustal architecture and structural evolution of the West African Craton are attributed to lateral shortening across the craton during the 2150–2070 Ma Eburnean Orogeny (e.g. Bonhomme, 1962; Eisenlohr & Hirdes, 1992; Feybesse, *et al.*, 2006). Following the extensive granitoid emplacement, an early phase of deformation is characterised by thrust-related tectonism, unit stacking and high-grade metamorphism attributed to crustal thickening (Allibone, *et al.*, 2002; Block, *et al.*, 2015; Block, *et al.*, 2016b; Feybesse, *et al.*, 2006; Galipp, *et al.*, 2003; Ganne, *et al.*, 2012; John, *et al.*, 1999; Milési, *et al.*, 1992). An alternative hypothesis suggests that the early crustal architecture defined dome and basin strain patterns, attributed to the diapiric uprising of granitoid bodies (Delor, *et al.*, 1995; Lompo, 2010; Vidal, *et al.*, 1996; Vidal, *et al.*, 2009). The widely recognised second phase of the Eburnean Orogeny (ca. 2110–2070 Ma) is characterised by the development of craton-scale strike-slip shear zones (Allibone, *et al.*, 2002; Feybesse, *et al.*, 2006; Jessell, *et al.*, 2012; Pouclet, *et al.*, 2006). Proponents of polycyclic orogenesis interpret high-grade terranes as older, deformed gneissic basement upon which discordant sequences of Birimian sediments were deposited during a crustal extension event which

were subsequently deformed during the Eburnean Orogeny (e.g. de Kock, *et al.*, 2012; Hein, 2010; Perrouty, *et al.*, 2012). In contrast, authors who invoke monocyclic orogenesis suggest that Birimian volcanic and sedimentary sequences formed pene-contemporaneously, underwent coeval deformation and metamorphism, and were subsequently juxtaposed during a single progressive deformation event during the Eburnean Orogeny (Block, *et al.*, 2015; Block, *et al.*, 2016b; Eisenlohr & Hirdes, 1992; Hirdes, *et al.*, 2007). Discrepancies in the geochemical, geochronological, isotopic evolution and tectono-thermal histories between terranes in the West African Craton have been used as evidence of suturing of independent cratonic fragments and collision orogenesis within the craton, (Block, *et al.*, 2016b; Parra-Avila, 2015; Parra-Avila, *et al.*, 2017).

5.3 Main lithologies and stratigraphy

The crustal architecture of southern Ghana and southeast Ivory Coast is defined by four parallel, elongate, narrow volcanic greenstone belts, the Kibi-Winneba, Ashanti, Sefwi and Bui belts (from southeast to northwest), separated by the Cape Coast, Kumasi-Afema and Sunyani-Comoé volcano-sedimentary basins (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1993; Jessell, *et al.*, 2012) (Fig. 5.2). The greenstone belts define the prevailing northeast-striking structural trend and are bounded by regional-to craton-scale shear zones. The stratigraphy of southern Ghana was recently re-characterised by Adadey, *et al.* (2009) and . The lower Sefwi Group, as defined by Perrouty, *et al.* (2012), comprises basalt and andesite sequences intercalated with minor volcaniclastic and sedimentary layers, with a minimum emplacement age of 2174 ± 2 Ma, interpreted from a syn-tectonic granitoid intrusion (U-Pb zircon age; Oberthür, *et al.*, 1998). The overlying Kumasi Group (2154 – 2125 Ma) is predominantly made of phyllites and volcaniclastic rocks, with syn-depositional andesitic volcanism dated in the Kumasi-Afema Domain at 2142 ± 24 Ma (Adadey, *et al.*, 2009). Detrital, fluvio-deltaic sedimentary rocks of the Tarkwa Group unconformably overlie the Sefwi Group volcanic rocks in the Ashanti belt, which were deposited between 2107 and 2097 Ma (Perrouty, *et al.*, 2012; Pigois, *et al.*, 2003). Existing geochronological data, including methodologies, for southern Ghana is summarised in Appendix D.



Adadey, et al. (2009); (3) Hirdes, et al. (1992); (4): Petersson, et al. (2017); (5) Hirdes and Davis (1998); (6): Petersson, et al. (2016); (7): Chapter 4 – this study; (8): Agyei Duodu, et al. (2009). Domain; HD = Hwidiem Domain; KJD = Kukuom-Juaboso Domain; SWD = Sefwi-Wiawso Domain; SZ = shear zone. Zircon age: (1) Hirdes, et al. (2007); (2): Fig. 5.2: Litho-structural map of the study area showing the location of samples discussed in this study and existing published geochronological data. CD = Chiraa

The study area focuses on plutonism and volcanism throughout the Sefwi Greenstone Belt and the volcano-sedimentary Sunyani-Comoé Domain (Fig. 5.2). These domains are bounded by the northeaststriking Sefwi Shear System (Jessell, *et al.*, 2012), including the Kenyase-Yamfo, the Ketesso and the Kumasi-Afema shear zones, from north to south. Results of the geochemical and geochronological analysis presented in this study are used to better understand the magmatic evolution of the study area, rather than redefining previously established stratigraphic column for the Ashanti Belt, located southeast of the Sefwi Belt.

5.4 Sampling and analytical methods

Twenty-nine fresh samples representative of the dominant lithological units, including nineteen granitoid samples and ten basaltic to rhyolitic volcanic and volcanoclastic rock samples, were collected from sub-domains within the Sefwi Greenstone Belt and the Sunyani-Comoé Domain for petrological and geochemical analysis. Samples were collected from the tectonic-metamorphic domains defined in Chapter 3, including the southern greenschist facies, volcano-plutonic Sefwi-Wiawso Domain, the central amphibolite facies Kukuom-Juaboso and Hwidiem domains, the Bechem Granitoid Domain and the north-eastern, migmatitic Chiraa Domain. Several additional samples were collected from volcanoclastic and plutonic units in the low-grade Sunyani-Comoé volcano-sedimentary domain. The locations of the samples are shown in Fig. 5.2 and coordinates are given in Table 5.1.

5.4.1 Whole rock geochemistry

Major oxide and trace element geochemistry was collected for all twenty-nine igneous samples. Powder concentrates for each sample were produced at the School of Earth, Atmosphere and Environment, Monash University. Major element oxide analysis was performed at the School of Earth Sciences, CODES, University of Tasmania (samples marked with asterisk in Table 5.1) and at the ALS Geochemical Laboratory, Brisbane, Australia, using X-ray Fluorescence (XRF) analysis according to ALS method ME-XRF26 (ALS, 2006a). Major element oxides analysed by XRF include SiO₂, TiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, CaO, Na₂O, K₂O, BaO, Cr₂O₃ and P₂O₅, with a lower detection limit of 0.01%. Loss on ignition was determined gravimetrically. Trace element concentrations, including rare earth elements, were obtained by Inductively Coupled Plasma-Atomic Emission Spectroscopy (ICP-AES) and Inductively Coupled Plasma–Mass Spectrometry (ICP-MS) at the ALS Geochemical Laboratory, Brisbane, Australia according to ALS geochemical procedure ME-MS61 (ALS, 2006b). Replicate analyses and Monash rhyodacite (DAND) and basalt (BNB) reference standards were included to

ensure reproducibility and accuracy. Major and trace element concentrations for all twenty-nine samples are presented in Table 5.1.

5.4.2 Zircon U-Pb dating

Four samples were selected for in-situ determination of the U-Pb age and Hf-isotope composition using Sensitive High-Resolution Ion Microprobe (SHRIMP) and Laser Ablation Multi Collector-Inductively Coupled Plasma Mass Spectrometer (LA-MC-ICP-MS) analysis, respectively.

5.4.2.1 Zircon separation

Zircon concentrates were obtained for five samples following the procedures described by Claoué-Long, et al. (1995). Zircon extraction followed the established techniques using magnetic fractionation using a hand magnet and a Franz Isodynamic magnetic separator, followed by heavy liquid (diiodomethane) separation. Zircons were hand-picked and mounted in epoxy resin discs, 5 mm in thickness and 25 mm in diameter. Samples were mounted with the following reference standards: M257 (561.3 Ma and 840 U ppm; Nasdala, et al., 2008), Temora II (416.8 Ma; Black, et al., 2004), OGC1 (3465 Ma; Stern, et al., 2009) and silicate glass NBS610. Reference standards were used for instrument set-up, common lead (²⁰⁴Pb) correction, Pb/U fractionation correction and to monitor instrument drift at the accuracy of results. Epoxy mounts were polished to expose a cross section through the internal structure of the grains. Back-scatter electron (BSE) and cathodoluminescence (CL) imaging of zircons was performed at the Monash Centre for Electron Microscopy (MCEM), Monash University using the JEOL JSM-7001F Field Emission Gun Scanning Electron Microscope (FEG SEM) and the FEI Nova NanoSEM 450 FEG SEM, respectively. Further BSE and CL imaging was performed at the Centre for Microscopy, Characterisation and Analysis at the University of Western Australia using the VEGA 3 Tescan SEM. BSE and CL images were used to identify internal structures, growth patterns and damage zones, and subsequently guide grain analysis.

5.4.2.2 U-Th-Pb SHRIMP II zircon dating

In-situ zircon U-Th-Pb analysis was conducted using a Sensitive High Resolution Ion microprobe (SHRIMP II) at the John de Laeter Centre for Isotope Research at Curtin University, Perth, Australia. Instrument conditions are detailed by Kennedy and De Laeter (1994) and De Laeter and Kennedy (1998). The analytical procedures for zircon dating with the SHRIMP ion microprobe are outlined in Claoué-Long, *et al.* (1995) and Williams (1998). The mass-filtered, 10keV primary O²⁻ beam is

focussed to an elliptical spot, ~25 μ m in diameter, with an ion current of mass-filtered, of 2.2 – 2.0 nA. For each spot, the dataset includes six cycles (scans) through the mass range of ¹⁹⁶Zr₂O⁺, ²⁰⁴ Pb⁺, ^{back}ground (204.1), ²⁰⁶Pb⁺, ²⁰⁷Pb⁺, ²⁰⁸Pb⁺, ²³⁸U⁺, ²⁴⁸ThO⁺ and ²⁵⁴UO⁺, as outlined by Nasdala, *et al.* (2008). Data reduction was performed using the MS Excel add-in programs, SQUID version 2.5 and ISOPLOT v3.0 (Ludwig, 2003), according to Compston, *et al.* (1984), Claoué-Long, *et al.* (1995) and Williams (1998). Applied common-Pb corrections were based on the Stacey and Kramers (1975) model for isotopic ratios of terrestrial lead. Analytical cut offs were applied prior to data interpretation in order to exclude zircons that had potentially undergone disruption of the U-Pb systematics. In this study, the reported U-Pb ages are derived from concordant or near-concordant grains with accepted discordance values between –5% and +15% in order to minimise the number of inversely concordant values. In addition, analyses that yielded U concentrations over 1000 ppm or included more than 1.0% of non-radiogenic Pb (²⁰⁴Pb) were rejected. All ages are reported as ²⁰⁷Pb/²⁰⁶Pb ages with 1 σ error.

5.4.3 In-situ Lu-Hf analyses

Lutetium-Hafnium (Lu-Hf) isotope analyses were performed on grains from samples dated using U-Pb zircon SHRIMP II method. Analyses were conducted using a Resonetics S-155-LR 193nm excimer laser ablation system attached to a Nu Plasma II multi-collector ICP-MS at the John de Laeter Centre, Curtin University. Sited adjacent to or "on-top-of" SHRIMP spots, circular laser spots with diameters of 33 µm were drilled with a laser pulse repetition of 5 Hz and an energy density of ~2 J/ cm² during 30 seconds of data acquisition, following an initial baseline period of 30 seconds. The isotopes ¹⁷³Yb, ¹⁷⁴(Yb, Hf), ¹⁷⁵Lu, ¹⁷⁶(Yb, Lu, Hf), ¹⁷⁷Hf, ¹⁷⁸Hf, and ¹⁷⁹Hf were measured on Faraday cups in a static sequence. Isobaric interference of ¹⁷⁶Yb was corrected according to the method of Nowell and Parrish (2001) whereby JMC475 Hf standard solutions were doped with Yb. From this, the true value of ¹⁷⁶Yb/¹⁷³Yb was derived using a regression, with ¹⁷⁶Yb subsequently subtracted from the 176 peak. Correction for interference of ¹⁷⁶Hf used the measured interference-free ¹⁷⁵Lu isotope and a value of 0.02653. Corrections were applied during each analysis after an inverse mass bias correction using the measured ¹⁷⁹Hf/¹⁷⁷Hf ratio, this allows for absolute differences between the Hf and Yb mass bias. The ¹⁷⁸Hf/¹⁷⁷Hf and ¹⁸⁰Hf/¹⁷⁷Hf stable isotope ratios were used to monitor data quality.

All data were adjusted relative to the primary reference zircon 91500. Analyses of reference zircons (MUN1, 3, 4,Fisher, *et al.*, 2011; R33, Fisher, *et al.*, 2014; GJ-1, Morel, *et al.*, 2008; Plešovice,

Sláma, *et al.*, 2008; 91500, Wiedenbeck, *et al.*, 1995; Mudtank, Woodhead & Hergt, 2005) were conducted every twenty spots throughout the session to verify accuracy and ensure reproducibility. Standard results are presented in Appendix E and comparison of measured standards to accepted values is shown in Appendix F.

The measured ¹⁷⁶Lu/¹⁷⁷Hf ratios for each of the zircons analysed were used to calculate initial ¹⁷⁶Hf/¹⁷⁷Hf ratios. There are a number of proposed decay constants for ¹⁷⁶Lu (e.g. Bizzarro *et al.*, 2003; Blichert-Toft *et al.*, 1997; Scherer *et al.*, 2001 ; Soderlund *et al.*, 2004). A decay constant (λ) for ¹⁷⁶Lu of 0.01867 for all Hf isotope calculations (Soderlund *et al.*, 2004, after Scherer *et al.*, 2001) is used in this study. EHf values are calculated relative to modern CHUR values, using the chondritic values of ¹⁷⁶Lu/¹⁷⁷Hf = 0.282785 and ¹⁷⁶Hf/¹⁷⁷Hf = 0.0336 (Bouvier, *et al.*, 2008, after Blichert-Toft and Albarède, 1997) are used for calculating EHf and model ages. Hf model ages (T_{DM-1}) are calculated for a linear 4.56 Ga depleted mantle model with modern compositions of ¹⁷⁶Lu/¹⁷⁷Hf = 0.0384 and ¹⁷⁶Hf/¹⁷⁷Hf= 0.28325 (Griffin, *et al.*, 2002). Two-stage model ages, which yield more robust model age estimates, are calculated with the same depleted mantle parameters with average crustal ¹⁷⁶Lu/¹⁷⁷Hf = 0.0150 (Condie, *et al.*, 2005) and a ¹⁷⁶Lu decay constant of 0.01867 (Söderlund, *et al.*, 2004 after Scherer *et al.*, 2001). Initial ¹⁷⁶Hf/¹⁷⁷Hf and $\varepsilon Hf_{(t)}$ were calculated for using individual zircon crystallisation ages to observe the array of zircon age relative to their Hf signature, as well as comparative $\varepsilon Hf_{(t)}$ plots of the study area relative to West African and global zircon Hf databases.

5.5 Results

5.5.1 Lithological and petrological descriptions

5.5.1.1 Granitoids

In order to simplify the presentation and the description of the lithologies sampled in this study, analysed granitoid samples have been grouped into four main lithologies, based on petrological composition and geochemical analysis, detailed in section 5.1 and 5.2, respectively. The four main groups are: i) tonalite, trondhjemite and granodiorite (TTG); ii) quartz diorite and diorite; iii) granite and quartz monzonite, and; iv) pyroxenite and gabbro norite.

Within the study area, there are two series of plagioclase-rich tonalite and trondhjemite plutons. The first group comprises large, composite, medium- to coarse-grained, leucocratic trondhjemite and tonalite plutons that contain biotite as the main ferro-magnesian mineral with subordinate to rare hornblende (Figs. 5.3a and 5.3b) and rare interstitial microcline. The most significant exposures are located in the Sefwi-Wiawso and Bechem Granitoid domains (Fig. 5.2). Accessory phases include euhedral magmatic epidote, euhedral titanite, apatite and zircon. Deformed plutons display biotite alignment and sub-grain rotation recrystallisation of large, subhedral to euhedral plagioclase phenocryst with irregular, bulging grain boundaries. Plagioclase shows zoning of some grains suggesting changes in anorthite content and is variably sericitized. Minor chlorite replacement of biotite indicates late, weak greenschist facies metamorphic overprint, increasingly pervasive proximal to pluton boundaries. The second group comprises smaller, elliptical to elongate, mesocratic granodiorite and trondhjemite plutons that are emplaced proximal to the Bibiani and Afema shear zones in the south of the Sefwi Belt and often contain small, round, mafic enclaves (Fig. 5.3c). The plutons are medium- to coarsegrained with an equigranular texture, containing hornblende as the primary ferro-magnesian mineral, subhedral to euhedral plagioclase and quartz. Minor biotite and microcline are often observed with accessory epidote, magnetite, apatite and hematite.

Quartz diorite and diorite plutons are medium- to coarse-grained and contain abundant hornblende phenocrysts and interlocking, subhedral to euhedral plagioclase often strongly altered by sericite, and minor quartz (Fig. 5.3d). Clinopyroxene phenocrysts often display extensive replacement rims of secondary hornblende and quartz. In the Hwidiem Domain, large irregular plutons are often weakly deformed displaying a foliation of aligned hornblende, potentially associated with retrograde metamorphism. Exposure of a mesocratic to melanocratic layered plutonic complex in the Bechem Granitoid Domain contains quartz diorite through to gabbro diorite, bounded to the south by the Ketesso High-Strain Zone. The plutons are variably deformed with foliations defined by metamorphic hornblende, seen wrapping around magmatic hornblende phenocrysts and are often structurally associated with strongly sheared metabasalts. Accessory minerals include magnetite, ilmenite, apatite and zircon with retrograde chlorite growth on hornblende rims and metamorphic epidote.

Three granite samples are grouped together based on their elevated potassium feldspar content, however, they each have distinct petrological assemblages and are presented together for comparative purposes.



Fig. 5.3. Representative samples: a) Medium-grained foliated biotite-hornblende trondhjemite; b) Typical trondhjemite assemblage including zoned, euhedral plagioclase (Pl) phenocrysts, aligned biotite (Bt), quartz (Qz) and magmatic, subhedral epidote (Ep); c) Outcrop of trondhjemite in Sefwi-Wiawso domain with mafic enclave containing retrograde chlorite and actinolite; d) Hand sample of mesocratic quartz diorite; e) Outcrop of coarse-grained potassic biotite-hornblende monzonite; f) Petrographic assemblage of potassic biotite-hornblende monazite with euhedral plagioclase displaying sericite alteration, interstitial microcline (Mc) and isolated, small hornblende (Hbl) and biotite grains; g)

Coarse grained to pegmatitic biotite-muscovite granite in metatectic paragneiss; h) Microcline-dominant leucogranite from the Sunyani-Comoé Domain, with plagioclase, quartz and muscovite (Ms).

Potassium-rich quartz monzonite plutons (Fig. 5.3e) occur as large (>20 km), irregular, weakly deformed composite plutons, predominantly restricted to within the Kukuom-Juaboso and Bechem Granitoid domains within the Sefwi Belt. They have a distinct radiometric signature associated with anomalously high K₂O content relative to U and Th (Chapter 3). The potassic plutons are leucocratic with a granular texture and contain zoned, euhedral plagioclase, microcline, quartz and biotite as the main ferromagnesian silicate (Figs. 5.3e; 5.3f), with minor hornblende and accessory magmatic epidote, titanite, ilmenite, magnetite and apatite. Granite samples from the Chiraa Domain are variably deformed, medium grained, irregular intrusions within high-grade metasedimentary rocks (Fig. 5.3g), comprising quartz, perthitic microcline and plagioclase. The metamorphic foliation within Chiraa Domain granites is defined by aligned biotite ± muscovite. Minor secondary chlorite is sometimes observed replacing biotite near the domain boundaries. Accessory minerals include ilmenite and zircon. Large, coarse-grained to megacrystic white-grey leucogranite plutons (>50 km) intrude the Sunyani-Comoé Domain, elongate parallel to the NE strike of the major crustal architecture. They contain quartz, muscovite, microcline with slight microperthitic texture, small euhedral plagioclase laths (Fig. 5.3h), and displaying weak sericite alteration. Weak deformation is indicated by irregular grain boundaries suggesting grain boundary migration.

Pyroxenite and gabbro plutons often form elongate (10 – 20 km) bodies, exposed in parts of the Bechem Granitoid Domain, as well as immediately south of the Hwidiem Shear Zone and in the southwestern extent of the Kukuom-Juaboso Domain (Fig. 5.2). Small (<10 km) elliptical gabbroic plutons are found in the study area, however, their emplacement timing and relation to the pyroxenites is unclear. The pyroxenite and gabbro plutons are mesocratic, coarse grained to megacrystic, equigranular plutons, ranging from undeformed to highly sheared (Fig. 5.4a). Petrographic assemblages include variable amounts of orthopyroxene, clinopyroxene, hornblende and plagioclase, depending on lithology and/or the presence of magmatic layering. Pyroxene grains are often extensively replaced by secondary hornblende and quartz, with late sericite alteration with some outcrops displaying abundant, centimetric plagioclase-rich veins (Fig. 5.4b). Accessory minerals include magnetite, ilmenite and secondary epidote and titanite.



Fig. 5.4. a) Hand sample of pyroxenite from the Bechem Granitoid Domain, containing clinopyroxene, minor orthopyroxene, both largely replaced by hornblende and quartz and minor plagioclase; b) highly altered and heavily veined outcrop of pyroxenite near the Ketesso Shear Zone; c) Typical metabasalt outcrop with chlorite-actinolite foliation; d) Vertical, foliated dacitic volcanoclasite (left) and strongly deformed volcano-sedimentary beds (right); e) Quartz-microcline-plagioclase-phyric rhyolite; f) Hornblende and plagioclase phyric dyke cross cutting folded amphibole-garnet paragneiss in the Kukuom-Juaboso domain.

5.5.1.2 Volcanic rocks

Volcanic rocks analysed in this study range from basaltic to rhyolitic in composition. These basalts, metabasalts and basaltic andesites contain magmatic and/or metamorphic hornblende (Fig. 5.4c). Foliations in the metabasalts and meta-basaltic andesites are defined by aligned hornblende in higher

grade samples, or chlorite and actinolite in lower grade samples. The deformed dacitic volcanoclastic rocks (Fig. 5.4d) contain relic plagioclase and quartz phenocrysts and exhibit chlorite-muscovite foliations with rare preservation of foliation aligned biotite porphyroblasts. A number of quartz-phyric pink rhyolite outcrops are documented, observed stratigraphically overlying garnet-hornblende paragneisses in the Kukuom-Juaboso domain (Fig. 5.4e). Late andesite dykes contain hornblende and zoned plagioclase phenocrysts, observed in the field as a cross-cutting dyke, intruding a folded amphibole-bearing para-amphibolite (Fig. 5.4f). Intermediate and felsic lavas commonly display sericite alteration, including growth of fine grained white mica.

5.5.2 Whole rock geochemistry

5.5.2.1 Samples

Samples were collected from most geological domains in the study area, including volcano-plutonic and volcanic layers in sedimentary domains, based on stratigraphic and structural relationships, and their potential significance for unravelling the early crustal evolution. A total of twenty-nine magmatic rocks were sampled for this study. They comprise nineteen plutonic samples, including four mafic to ultramafic plutons, and ten volcanic samples. Sample locations are shown in Fig. 5.2 and the results are presented in Table 5.1.

5.5.2.2 Plutonic rocks

Sampled plutonic rocks record a wide range of silica contents. Plutons in the granite and TTG groups have high silica contents of 70–76 wt% SiO₂, with the exception of the quartz monzonite sample with 68 wt% SiO₂ (Fig. 5.5a). Diorite and quartz diorite rocks range from 55 to 61 wt% SiO₂ (Fig. 5.5a), whilst pyroxenites and gabbros contain < 52 wt% SiO₂ (Table 5.1 – not shown in Fig. 5.5). Excluding the mafic and ultramafic suites, the silicic, plagioclase-rich granitoids are plotted on an An-Ab-Kfs normative ternary diagram (fields from Barker, 1979; O'Connor, 1965), distinguishing tonalite, trondhjemite, granodiorite and granite rocks based on normative CIPW norms, specifically normative feldspar composition. Six samples plot in the trondhjemite field, six in the tonalite field and three in the granite field (Fig. 5.5b). Interestingly, no samples plot within the granodiorite field.

Table 5.1: Whole rock	major (wt%)) and trace e	lement (ppm) compositio	n of magmai	tic rocks from	n SW Ghana	SD 072	SD 07(SD 400
Sample	3B 005	Leuco-	SB 010"	SB 023	SB 040*	5B 044	SB 045	SB 072	SB 070	Trachy-
Lithology	Diorite	granite	Dacite	Granite	Gabbro	Rhyolite	Tonalite	Trondjh.	Rhyolite	dacite
Latitude	6.7468451	6.9112397	6.5552958	6.7104697	6.7435252	6.4807334	6.4808303	6.2387206	6.3766174	6.8753326
Longitude	-2.490747	-2.625645	-2.616664	-2.680397	-2.482046	-2.331715	-2.357751	-2.486864	-2.679099	-2.468147
Age (Ma)				2136						
Error (1σ)				7						
SiO	56.61	73.00	64 33	67.07	52 78	76 30	71.16	72.01	72.85	67.74
510 ₂	0.01	13.99	04.55	07.97	0.70	0.16	/1.10	0.25	12.05	07.74
1102	0.85	0.08	0.85	0.10	0.79	0.10	0.32	0.25	0.05	0.43
Al_2O_3	15.16	14.92	13.24	17.12	14.77	11.66	13.52	15.22	15.34	17.80
FeO _T ^a	6.94	0.50	8.24	1.56	8.79	1.95	3.32	2.21	0.68	2.44
MnO	0.12	0.01	0.15	0.04	0.14	0.04	0.07	0.03	0.01	0.01
MgO C=O	5.63	0.17	1.12	0.45	6.95	0.37	0.96	0.60	0.11	0.67
CaO N= O	0.07	0.74	3.52	2.45	8.06	1.19	2.95	2.72	0.24	0.21
Na ₂ O	3.40	4.44	4.21	5.28	3.64	3.84	3.78	4.95	5.81	5.30
K ₂ O	2.01	4.47	1.39	4.01	1.17	2.30	1.82	1.09	4.23	3.38
P_2O_5	0.34	0.09	0.29	0.05	0.15	0.01	0.07	0.07	0.01	0.12
LOI ^D	1.75	0.64	1.73	0.70	1.84	1.69	1.19	1.02	0.43	1.81
Total	100.55	100.20	89.10	100.20	88.44	99.98	99.70	100.60	100.20	100.40
K ₂ O/Na ₂ O	0.59	1.01	0.33	0.76	0.32	0.60	0.48	0.22	0.73	0.64
Mg# ^c	59.12	37.74	19.50	33.96	58.50	25.27	34.01	32.61	22.38	32.86
$FeO_T + MgO$	12.57	0.67	9.36	2.01	15.74	2.32	4.28	2.81	0.79	3.11
A/NK	1.95	1.23	1.57	1.31	2.04	1.32	1.65	1.63	1.09	1.44
A/CNK	0.76	1.11	0.89	0.98	0.68	1.06	1.00	1.07	1.05	1.40
Trace elements (ppm)										
Cs	1.54	23.8	0.49	1.09	0.91	0.71	0.63	0.59	0.45	3.72
Rb	69.8	260.	39.8	122.	23.4	62.7	49.4	20.3	81.	124.
Ba	610.	460.	690	890.	290	870.	690.	630.	1900.	1040.
Nb	8.2	5.2	6	2.7	6.5	8.8	8.2	5.1	1.5	4.4
Sr	661.	181.	289	884.	460	83.5	143.	479.	1090.	348.
Zr	120.	57.6	165.5	43.4	53.5	206.	117.5	103.	68.2	169.5
Y	23.5	2.8	41.1	9.6	20.5	25.6	31.2	3.9	3.1	5.9
La	30.5	12.	18.4	12.5	10.7	33.9 65.6	26.4	/.6	1.5	25.3
Dr.	0.63	20.5	43.8	2.88	24.9	03.0 7.01	52.1 5.74	14.5	2.02	57.5
Nd	39.3	11.3	26.9	12.88	14.8	25.8	21.7	5.9	1.8	20.2
Sm	7.6	2.35	6.29	2.51	3.44	4.82	4.2	0.96	0.41	3.22
Eu	1.98	0.53	1.92	0.8	1.07	0.83	0.91	0.4	0.12	0.81
Gd	5.82	1.51	7	2.46	3.64	4.21	4.24	0.65	0.46	2.02
Tb	0.79	0.16	1.12	0.32	0.58	0.64	0.71	0.1	0.07	0.23
Dy	4.47	0.65	7.26	1.78	3.72	4.09	4.78	0.64	0.44	1.19
Но	0.84	0.09	1.54	0.33	0.79	0.84	1.02	0.14	0.1	0.21
Er	2.32	0.2	4.5	0.9	2.34	2.65	3.09	0.42	0.29	0.59
1m Vh	0.33	0.03	0.66	0.13	0.33	0.42	0.48	0.07	0.05	0.09
iu Iu	0.32	0.17	4.20	0.82	0.33	2.9	5.15 0.51	0.47	0.55	0.37
Hf	3.2	2.1	4.6	1.5	1.6	6.2	3.7	2.6	2.6	4 9
Та	0.59	1.	0.46	0.38	0.51	0.75	0.78	0.66	0.22	0.38
Th	5.1	4.2	2.4	0.5	1.8	6.3	5.7	2.3	0.3	7.1
U	1.9	7.2	0.9	0.6	0.9	1.5	2.	0.8	1.	3.2
Pb	7.2	31.2	4.4	11.2	3.7	7.8	8.3	6.7	18.5	17.8
Sc	20	1	21	3	35	6	8	3	1	6
V	153	5	28	22	211	4	23	19	6	44
Cr	219	3	1	4	217	1	11	<1	3	26
C0	58	85	89	85	77	60	59	90	73	38
INI Ga	122	2	1 21 4	د د ۸د	81 1775	< 14.4	14	10.05	2	19 200
Са Р	1520	24.1 410	21.0 1340	24.3 250	17.75	14.0	340	19.95	23.7	20.8 540
Ba/Rb	8 74	1 77	17 340	230 730	12 30	13.88	13 97	31.03	23.46	540 8 30
Sr/Y	28.13	64.64	7.03	92.08	22.44	3.26	4.58	122.82	351.61	58.98
Nb/Ta	13.90	5.20	13.04	7.11	12.75	11.73	10.51	7.73	6.82	11.58
Lu/Hf	0.100	0.014	0.141	0.087	0.206	0.079	0.138	0.035	0.023	0.018
(La/Yb) _N	9.77	47.95	2.92	10.36	3.30	7.94	5.73	10.98	2.91	30.15
Eu _N /Eu*	0.91	0.86	0.88	0.98	0.92	0.56	0.66	1.55	0.84	0.97

^a Total Fe and FeO

^b Loss on Ignition

^c 100*(Molar Mg/(Mg + FeII))

Table 5.1: cont.										
Sample	SB 090*	SB 092	SB 154*	SB 164*	SB 173	SB 178b	SB 191	SB 204	SB 206	SB 217
Lithology Latitude Longitude Age (Ma) Error (1o) Maior alements (w?%)	Andesite 6.9724842 -2.446067	Qz. diorite 6.9716938 -2.377918 2144 10	Pyrox. 6.403997 -2.79812	Basalt 6.4427289 -2.78907	Dacite 7.3801879 -2.290199 2158 6	Trondjh. 7.4840981 -2.166146	Gabbro 7.1636225 -2.077912	Rhyolite 6.1971109 -2.401886	Qz. diorite 6.1975668 -2.438266	Trondjh. 6.0576222 -2.57115
SiO-	63.01	61.83	51 73	48 72	68.81	74.68	48.61	77 30	56.15	74.96
TiO.	0.50	01.05	0.51	1 30	0.64	0.07	1.84	0.15	0.85	0.25
Al-O-	16.94	16.00	4.26	17.01	14 38	15.64	14 58	11.46	15.48	12.20
Fac b	6.20	10.00	10.21	12.15	5.62	0.55	11.50	2.02	0.20	2.20
MpO	0.20	4.28	0.23	0.10	5.05 0.15	0.55	0.20	2.02	9.38	2.04
ΜσΟ	2.82	3.08	15 30	6.63	1.64	0.01	6.81	0.03	3 54	0.03
CaO	1.84	4.70	14.32	10.45	0.44	1.50	10.95	0.80	7.33	1.32
Na ₂ O	2.45	4.93	0.75	0.75	4.09	4.36	2.28	4.27	3.56	4.42
K ₂ O	2.25	2.05	0.15	0.12	1.54	1.99	0.41	2.80	0.84	2.49
P_2O_5	0.09	0.22	0.10	0.12	0.14	0.02	0.22	0.01	0.13	0.03
LOI ^c	2.79	1.47	1.29	0.36	2.11	0.90	0.90	0.46	1.27	0.67
Total	90.11	99.96	87.36	85.38	100.45	100.05	100.20	99.84	100.00	99.72
K ₂ O/Na ₂ O	0.92	0.42	0.20	0.16	0.38	0.46	0.18	0.66	0.24	0.56
$Mg\#^d$	44.77	56.19	72.57	47.33	34.18	36.84	51.16	1.73	40.22	13.44
FeO _T + MgO	9.03	7.36	25.62	19.78	7.27	0.73	18.40	2.04	12.92	2.87
A/NK	2.61	1.55	3.05	12.47	1.71	1.68	3.48	1.14	2.29	1.22
A/CNK	1.72	0.85	0.16	0.84	1.56	1.30	0.60	1.00	0.77	0.99
Trace elements (ppm)										
Cs	2.96	0.94	0.05	0.12	3.3	0.81	0.92	0.32	0.37	1.5
Rb	78.1	37.1	1.3	0.5	63.1	49.7	10.8	66.6	9.9	71.3
Ba	730	890	40	10	510	440	170	960	380	700
Sr	373	1020	1.5	3.0	320	2.9 352	345	63.1	230	59.8
Zr	175	65.3	34.9	21.5	131.5	61.5	143	211	117.5	188.5
Y	19.8	11.1	18	19	19	0.9	28.2	103.5	20.8	39.3
La	17.9	32.7	9.1	4	26.3	4.9	17.9	67	10.5	27.1
Ce	177	69.7	25.8	11.5	56.5	10.25	39.7	128.5	23.2	61.2
Pr Nd	6.59 24.0	8.06	4.51	1.9	6.53 25.4	1.2	5.11	14.8	2.78	6.14 23 7
Sm	51	4 94	5 23	2.83	4.6	1.04	4 78	12.9	2.58	4 81
Eu	1.41	1.4	1.31	1.14	1.27	0.37	1.69	2.31	0.98	0.95
Gd	4.17	3.44	4.68	3.47	3.8	0.62	5.25	14.45	3	5.01
Tb	0.84	0.42	0.64	0.57	0.55	0.06	0.84	2.39	0.52	0.88
Dy	5.72	2.24	3.57	3.71	3.34	0.21	5.24	15.95	3.55	6.08
H0 Fr	1.17	1.09	0.08	2.12	0.09	0.03	1.07	5.45 10.15	0.79	1.55
Tm	0.59	0.16	0.24	0.3	0.29	0.05	0.42	1.52	0.37	0.64
Yb	4.27	0.97	1.53	1.85	1.88	0.05	2.58	9.91	2.49	4.23
Lu	0.63	0.16	0.22	0.27	0.29	0.01	0.39	1.51	0.4	0.66
Hf	4.7	1.7	1.3	0.9	3.8	2.6	3.7	7.4	3.1	5.8
la Th	0.54	0.46	0.08	0.3	0.45	0.54	0.92	1.35	0.43	1.19
U	2.2	4.7	0.2	0.3	4.0	2.3	0.4	2.3	0.6	2.3
Pb	10.2	8.5	<0.1	1.9	8.6	15.1	3.3	5.2	2.7	6.9
Sc	16	9	78	27	14	1	44	3	22	6
V	86	88	234	256	89	5	305	1	158	7
Cr	57	97	574	6	93	2	133	4	9	3
C0 Ni	43 1	52	02	117	46	97 2	80 86	119	67 27	142
Ga	22	21	93 937	2.2	50 18 1	23.9	20.8	5 18 7	1915	15.85
P	400	1020	470	530	620	70	990	50	610	160
Ba/Rb	9.35	23.99	30.77	20.00	8.08	8.85	15.74	14.41	38.38	9.82
Sr/Y	18.84	91.89	6.03	17.21	16.84	391.11	12.23	0.61	11.06	1.52
Nb/Ta	14.81	16.30	16.25	12.00	13.56	5.37	17.07	12.00	13.95	10.00
Lu/Ht (Lo/Vb)	0.134	0.094	0.169	0.300	0.076	0.004	0.105	0.204	0.129	0.114
(La/10) _N Eu./Eu*	2.85	22.90	4.04	1.47	9.50	00.5/	4./1	4.59	2.86	4.35
Lu _N /Lu	0.95	1.04	0.81	1.11	0.95	1.41	1.05	0.32	1.08	0.39

Table 5.1: cont.									
Sample	SB 244	SB 248	SB 252	SB 347 Basaltic	SB 370	SB 406	SB 408	SB 411	SB 464B
Lithology	Qz. diorite	Trondjh.	Trondjh.	andesite	Granite	Trondjh.	Pyrox.	Diorite	Dacite
Latitude	6.1955922	6.2680786	5.9885924	6.0268026	7.6065118	7.0845057	7.0922144	7.0320639	6.8012179
Longitude	-2.427349	-2.481873	-2.634561	-2.873853	-2.134411	-1.981516	-1.946243	-1.920252	-2.388421
Age (Ma)		2155							
Error (1σ)		5							
Major elements (wt%)									
SiO ₂	60.74	71.15	70.78	56.29	73.41	71.06	47.32	59.71	68.17
TiO ₂	0.68	0.23	0.42	0.73	0.12	0.24	0.58	0.62	0.28
Al_2O_3	15.51	15.73	13.90	15.46	15.30	15.44	5.97	16.64	15.50
FeO _T ^b	6.53	1.68	4.11	7.70	0.63	1.68	9.83	6.03	2.50
MnO	0.11	0.02	0.08	0.12	0.01	0.02	0.19	0.10	0.04
MgO	2.54	0.66	0.54	5.11	0.21	0.55	15.40	3.29	0.82
CaO	6.01	2.74	2.64	8.45	0.85	2.31	14.30	5.83	3.47
Na ₂ O	3.63	5.52	4.43	2.95	4.56	5.47	0.81	3.94	4.89
K ₂ O	1.35	1.24	1.84	0.37	4.23	1.92	0.19	0.81	2.68
P_2O_5	0.24	0.07	0.10	0.13	0.07	0.06	0.04	0.16	0.11
I OI ^c	1.70	0.57	0.00	1.01	0.62	0.41	2.66	1.50	0.76
LOI Total	100.05	0.37	100.45	1.91	100.05	0.41	00.82	00.50	0.70
K O/Na O	0.37	99.97	0.42	0.13	0.03	99.32	0.23	99.50	99.75
$K_2 O/Ma_2 O$	0.57	0.22	0.42	0.15	0.93	0.55	0.25	0.21	0.55
Mg#"	40.95	41.19	18.98	54.19	37.27	36.85	73.63	49.30	36.90
$FeO_T + MgO$	9.07	2.34	4.65	12.81	0.84	2.23	25.23	9.32	3.32
A/NK	2.09	1.51	1.50	2.94	1.27	1.39	3.88	2.26	1.42
A/CNK	0.84	1.02	0.99	0.75	1.12	1.01	0.22	0.93	0.90
Trace elements (ppm)									
Cs	0.43	0.98	0.29	0.14	26.2	0.68	<0.05	1 91	1 46
Rb	16.7	34.5	47	3 4	327	38.5	-0.05	1.51	87.8
Ba	610	430	710	130	500	590	40	510	780
Nb	5.4	3.1	10.5	4.7	7.7	2	1.1	5.4	4.6
Sr	345	579	155.5	179.5	187.5	600	94.7	350	985
Zr	119	58.4	162	103.5	52	51.1	18.8	55.6	99.5
Y	16.5	2	37	16.5	3.7	5.1	13.3	8.8	6.4
La	20.8	15.4	30.8	10	12.5	12.1	2.2	9.3	8.9
Ce	46.6	21.6	62.2	21.3	29.5	27.1	6.11	20.1	18.35
Pr	5.43	2.66	6.91	2.49	3.24	3.06	1.01	2.23	2.53
Nd	21.4	8.8	26.8	10.1	12.1	11.9	5.5	1 70	10.3
Sm Fu	3.70	1.08	5.15	2.19	2.83	2.55	1.7	1.79	2.24
Gd	3 23	0.43	5.17	2.45	2.01	1.71	2.15	1.78	1.71
Th	0.47	0.00	0.85	0.41	0.23	0.22	0.34	0.27	0.22
Dv	2.94	0.33	5.65	2.76	0.96	1.13	2.26	1.66	1.16
Но	0.61	0.06	1.23	0.59	0.13	0.2	0.47	0.34	0.21
Er	1.78	0.17	3.75	1.78	0.27	0.51	1.33	0.99	0.56
Tm	0.27	0.02	0.57	0.27	0.04	0.07	0.19	0.14	0.08
Yb	1.76	0.16	3.7	1.75	0.2	0.44	1.19	0.97	0.48
Lu	0.28	0.03	0.59	0.27	0.03	0.07	0.19	0.15	0.07
Hf	3.1	1.5	5	2.7	2	1.5	0.8	1.7	2.9
Ta	0.48	0.47	0.91	0.38	3.77	0.32	0.08	0.49	0.36
ln U	3	1.5	4.7	1	5.2	1.3	<0.2	1.6	1.8
U Ph	58	0.5	1.9	0.4	14.1	0.0	0.1	0.8	0.8
Sc	15	0.5	11	23	10.5	3	64	 14	5
V	114	20	13	142	7	22	206	102	49
Cr	24	12	3	119	3	5	1240	53	8
Со	53	128	96	62	79	94	73	51	51
Ni	14	6	1	110	2	5	208	47	5
Ga	16.95	20.8	18.35	15.95	22.1	19.95	8.62	18.7	21.8
Р	1080	320	440	560	350	300	190	750	530
Ba/Rb	36.53	12.46	15.11	38.24	1.53	15.32	40.00	39.84	8.88
Sr/Y	20.91	289.50	4.20	10.88	50.68	117.65	7.12	39.77	153.91
Nb/Ta	11.25	6.60	11.54	12.37	2.04	6.25	13.75	11.02	12.78
Lu/Ht	0.090	0.020	0.118	0.100	0.015	0.047	0.238	0.088	0.024
$(La/Yb)_{N}$	8.03	65.39	5.65	3.88	42.46	18.68	1.26	6.51	12.60
Eu _N /Eu*	0.99	1.63	0.67	1.02	0.68	1.15	0.99	1.20	1.05

Quartz-diorites, diorites and most granites plot along the calc-alkaline differentiation trend in the Na-K-Ca ternary diagram. The more sodic trondhjemite and tonalite samples, as well as the quartz monzonite, plot in the Archaean trondhjemite field (Fig. 5.5c). In the Harker diagrams (Fig. 5.5a), the SiO_2 vs. K_2O plot shows that the trondhjemite plutons display both tholeiitic (TH) and calc-alkaline (CA) affinities, whilst quartz-diorites and diorites plot primarily in the CA series, with one sample falling in the high-potassium calc-alkaline (HKCA) series. The quartz monzonite and granite samples all plot in the HKCA series. Overall, oxide and silica relationships in the Harker diagrams show approximately linear differentiation trends, with increasing SiO_2 associated with decreasing MgO, CaO, TiO₂ and FeO. Poor correlation, however, is observed between Al_2O_3 , Na_2O and K_2O in relation to increasing SiO₂, consistent with sericite alteration.



Fig. 5.5. Major element geochemistry of plutonic rocks of the study area, presented in (a) Harker diagrams (excluding

pyroxenites and gabbros), (b) normative feldspar triangle (excluding pyroxenite and gabbro), (c) molecular Na-K-Ca ternary diagram and (d) in an A/CNK vs A/NK diagram. Symbol legend (top right) represents the lithological groups based on identification of major mineral assemblages and the classification in the feldspar diagram.

For the tonalite-trondhjemite plutons, K_2O/Na_2O values range from 0.22 to 0.48, with one sample (SB045) plotting on the tonalite-granodiorite boundary with a K_2O/Na_2O ratio of 0.56. Quartz diorite plutons display similar ranges (0.21 – 0.42) with a meta-gabbro recording an elevated K_2O/Na_2O of 0.59, likely as a product of metamorphic processes. The granite and quartz monzonite plutons record significantly higher ratios of 0.76–1.01 relative to both the tonalite-trondhjemite and quart-diorite plutons.

Tonalite-trondhjemite plutons have Mg# (100*(Molar Mg/(Mg/Fe²⁺)) of 13–31, with lower values often representing hornblende-dominant granitoids, with concentration of mafic oxides ranging between 0.73 and 4.65. By comparison granite and quartz monzonite samples have intermediate Mg# (34–37) and low mafic oxide concentrations (MgO + FeO_T = 0.67–2.01). Quartz diorites and diorites display Mg# an order of magnitude higher (40–59) and mafic oxide concentrations of 7.36–12.92 wt%. Pyroxenite and gabbro samples have low K₂O/Na₂O (0.18–0.32), and elevated Mg# (51–73) and mafic oxide concentration (MgO + FeO_T = 18.4–25.6 wt%). Tonalites and trondhjemites are slightly metaluminous to slightly peraluminous (A/CNK = 0.99–1.07; A/CNK: molar ratio Al₂O₃/[CaO + Na₂O + K₂O] and A/NK = 1.22–1.65; A/NK: molar ratio Al₂O₃/[Na₂O + K₂O]) with one anomalously peraluminous sample (A/CNK = 1.30; A/NK = 1.68). Granites are peraluminous (A/CNK = 1.31; A/NK = 0.98). Quartz-diorite, pyroxenite and gabbro samples are strongly metaluminous, yielding ratios of A/CNK = 0.76–0.93; A/NK = 1.55–2.26 and A/CNK = 0.16–0.68; A/NK = 2.04–3.88, respectively.

Trace element geochemistry of the plutonic samples are presented in primitive mantlenormalised multi-element and chondrite-normalised rare earth element (REE) spidergrams (Fig. 5.6). All samples show strongly negative anomalies for high field strength elements (HFSE) (Ta, Nb, Ti) with the exception of one meta-gabbro sample (SB191). Most samples also display a positive Pb anomaly, with the exception of granite samples SB010 and SB370. Pyroxenites and gabbros are weakly fractionated with low LREE contents (La_N = 9.28–45.15) and high HREE and Y contents (Yb_N = 7.39– 16.02), thus yielding low LREE/HREE ratios (La_N/Yb_N = 1.25–4.04). They display slightly negative to no Eu anomalies (Eu_N/Eu^{*} = 081–1.03; Eu^{*} = [Sm_N × Gd_N]^{1/2}), and extremely low enrichment in large



5.6. Primitive-mantle normalised multi-element (McDonough, et al., 1992) and chondrite-normalised REE (McDonough & Sun, 1995) spidergrams.

ion lithophile elements (LILE) ($Rb_N/La_N = 0.01-0.22$), associated with moderately negative Rb, weakly positive Ba and mixed Sr anomalies (Sr = 94–460 ppm). Furthermore, only pyroxenites and gabbros samples display positive V and Sc anomalies with the exception of meta-gabbro sample SB040, which

has a slightly negative Sc anomaly potentially indicating a dioritic protolith rather than a gabbroic protolith. Quartz-diorite and diorite samples reveal intermediate REE fractionation patterns (La_N / Yb_N = 2.86–22.90), with moderate to high LREE contents (La_N = 39.24–137.97) and high HREE and Y content (Yb_N = 6.02–15.46). They display slightly negative to slightly positive Eu anomalies (EuN/ Eu* = 0.91–1.20), and minimal LILE enrichment relative to LREE (Rb_N/La_N = 0.08–0.24), much like the pyroxenes and gabbros.

Both the multi-element diagram and REE diagram reveal two distinct tonalite-trondhjemite lithological groups, corresponding with the biotite-bearing (black triangles, Fig. 5.6) versus the hornblende-dominant (red triangles, Fig. 5.6) tonalite-trondhjemite plutons, as described in Section 5.1. The biotite-rich samples have moderate to high LREE values ($La_N = 20.67-64.97$) and low HREE - Y concentrations (Yb_N = 0.31-2.92) contents, resulting in a steep, more fractionated REE spectra $(La_N/Yb_N = 10.98-66.57)$. The hornblende-rich samples have flat, moderate fractionation REE spectra, corresponding with both high LREE concentration ($La_N = 111.39 - 129.95$) and high HREE – Y concentration (Yb_N 19.44 – 26.27), yielding a moderately fractionated REE pattern (La_N/Yb_N=4.35– 5.72). Additionally, the two groups have contrasting Sr and Eu anomalies, with the more heavily fractionated, biotite dominant group displaying strong positive Sr anomalies (Sr = 352-600ppm), positive Eu anomalies (Eu_N/Eu* = 1.14-1.62), as well as low to moderate LILE enrichment relative to LREE ($Rb_N/La_N = 0.23-1.04$), whilst the less fractionated, hornblende dominant group displays, negative Sr anomalies (Sr = 59–156 ppm), negative Eu anomalies (Eu_N/Eu* = 0.59–0.67) and low LILE enrichment relative to LREE ($Rb_N/La_N = 0.15-0.27$), reflecting the high LREE concentrations. Contrasting trace element chemistry suggests variation in the petrogenetic history of the two tonalitetrondhjemite groups.

Granites samples are characterised by moderate LREE content ($La_N = 50.63-52.74$), moderate to high HREE – Y contents ($Yb_N = 1.06-1.24$) and moderate to strong REE fractionation patterns ($La_N/Yb_N = 42.45-47.95$), with the higher HREE values yielded by the quartz monzonite, indicating weaker REE fractionation ($La_N = 52.75$, $Yb_N = 5.09$, $La_N/Yb_N = 10.35$). Relative to all other samples, the granites display strong LILE enrichment ($Rb_N/La_N = 1.01 - 2.69$) as well as weakly negative to negative Eu anomalies ($Eu_N/Eu^* = 0.67 - 0.98$).

Samples from volcanic units across the study area are plotted on the TAS diagram (Le Maitre, *et al.*, 1989) displaying a range of compositions from basalt to rhyolite (Fig. 5.7a). Nine of thet en samples plot in the sub-alkaline field, whilst a quartz-feldspar-phyric rhyolite sample (SB076) is slightly alkaline. In the AFM diagram , all samples plot (Fig. 5.7b) in the calc-alkaline field except for a metabasalt (SB164) and one meta-dacite (SB016), both from the Kukuom-Juaboso Domain within the Sefwi Belt. All dacitic samples plot close to the calc-alkaline field boundary, indicating more Fe-rich compositions.

Meta-basalt sample SB164 displays positive Pb and Sr anomalies, a negative Zr anomaly, and a relatively flat trace element spectra. Both meta-basalt and basaltic andesite samples have similar REE patterns characterised by moderate LREE ($La_N = 16.88-42.19$) and HREE- Y ($Yb_N = 10.87-11.49$) contents, and low REE fractionation ($La_N/Yb_N = 1.47 - 3.88$). The basalt-andesite displays weak LILE enrichment with weakly positive Ba, U and La anomalies, as well as a positive Pb anomaly. Overall the group shows very weak LILE enrichment ($Rb_N = 0.21-1.48$; $Rb_N/La_N = 0.01-0.04$) and neither sample displays a Eu anomaly ($Eu_N/Eu^* = 1.02 - 1.11$).

Dacites display more fractionated REE patterns, with moderate to high LREE content ($La_N = 37.55-110.97$) and moderate to high HREE – Y content ($Yb_N = 2.98 - 26.58$), and $La_N/Yb_N = 2.86-30.15$. Dacites display LILE enrichment ($Rb_N = 17.34 - 53.91$; $Rb_N/La_N = 0.22-1.02$), strong negative HFSE anomalies (Nb, Ta, Ti) and no Sr anomalies, with the exception of a plagioclase-phyric dacite sample (SB464B), which shows a strong positive Sr anomaly. Foliated dacitic volcanoclastic sample SB090 shows unique trace element patterns, displaying a distinct positive Ce anomaly and negative Y anomaly. Eu anomalies in dacite samples are slightly negative to neutral ($Eu_N/Eu^* = 0.88 - 1.05$).

Rhyolitic samples display heterogeneous trace element patterns in multi-element diagrams, although each have negative Nb, Tb and Ti anomalies and negative Eu anomalies ($Eu_N/Eu^* = 0.52-0.85$). REE spectra for each of the rhyolites show similarly moderate REE fractionation patterns ($La_N/Yb_N = 2.91-7.94$), however, quartz-feldspar-phyric rhyolite sample (SB076) contrasts all other felsic lavas. It displays extreme LILE enrichment ($Rb_N = 35.22$; $Rb_N/La_N = 5.56$) with strong positive Pb, Ba, K and Sr (1090 ppm) anomalies, depletion in MREE and a weakly negative Eu anomalies (EuN/Eu* = 0.85) and a REE pattern less than an order of magnitude greater than chondrite values (Yb_N = 2.17).



Fig. 5.7. (a) TAS diagram (Le Maitre, et al., 1989) displaying volcanic samples from across the field area, ranging from basalt (circle) to basaltic andesite (diamond) to dacite (triangle) and rhyolite (square). (P-B) Picro-basalt; (B-A) Basaltic andesite; (A) Andesite; (T-A) Trachyandesite; (B-T-A) Basaltic trachyandesite; (T-B) Trachybasalt; (T/B) Tephrite/ Basanite; (P-T) Phonotephrite; (T-P) Tephriphonolite. (b) AFM ternary diagram (Irvine & Baragar, 1971; Kuno, 1968) where A is $Na_{,O} = K_{,O}$, F is FeO_T and M is MgO. (c, e, g) Primitive-mantle normalised trace element (McDonough, et

Conversely, sample SB044 and SB244 are enriched in most incompatible elements and have negative Sr (63–83 ppm) and Eu ($Eu_N/Eu^* = 0.51-0.56$), with sample SB044 displaying extreme enrichment in both HREE and LREE ($La_N = 282$; Yb_N = 61.55).

5.5.3 Geochronology

5.5.3 Results of zircon U-Pb and Lu-Hf analysis

In-situ zircon SHRIMP U-Pb analyses were performed on four samples, including three plutonic samples and one volcanic sample. This was complemented by Lu-Hf isotope analysis by LA-MC-ICP-MS, both performed at Curtin University. An additional detrital zircon sample was selected for in-situ U-Pb analysis by LA-ICP-MS at Macquarie University. Back-Scattered Electron (BSE) and cathodoluminescence (CL) images of representative zircons grains analysed in each sample. CL images are presented in Figure 5.8. The results of U-Pb and Lu-Hf analyses are shown in Tables 5.2 and 5.3 and Figures 5.9, 5.10 and 5.11.

5.5.3.1 Foliated biotite trondhjemite (sample SB248)

Zircons from biotite trondhjemite sample SB248 are commonly euhedral and prismatic, ranging in shape from stubby to elongate. Grains are approximately 100 to 250 μ m in length and 50–100 μ m in width. Smaller prismatic grains are pale coloured and relatively free of fractures whilst larger grains show brown colouration and more extensive fracture networks. Cathodoluminescence images reveal fine oscillatory zoning with no clear evidence of xenocrystic cores (Fig. 5.8a). Rare examples of thin (<10 μ m) metamict rims were visible in CL images.



Fig. 5.8. Cathodoluminescence (CL) images of representative zircon grains from analysed samples showing SHRIMP spot location (red dashed circles).

Twenty-six analyses were performed on twenty-five zircons. Six analyses were rejected: one because it has a discordancy greater than +10%; one analysis that had a ²⁰⁴Pb corr % ²⁰⁶Pb greater than 1%; three that had 1 σ greater than 25 Ma and one analysis which had U content over 1000 ppm. Twenty accepted analyses from sample SB248 yielded a single age population and a weighted mean average ²⁰⁷Pb/²⁰⁶Pb age of 2155 ± 5 Ma (MSWD of 1.15) interpreted to be the crystallization age due to the smaller error

relative to the intercept age (Fig. 5.9a). Concordia diagrams indicate a similar upper intercept age of 2155 ± 14 (MSWD of 0.55) (Fig. 5.9b), with the Discordia line indicating minor lead loss.

Twenty Hf analyses in dated grains yielded ¹⁷⁶Lu/¹⁷⁷Hf of < 0.0007, ¹⁷⁶Yb/¹⁷⁷Hf of < 0.02 and ¹⁷⁶Hf/¹⁷⁷Hf range from 0.281543 to 0.281630 ± 0.000034–0.000067 (2SE). Individual Pb-Pb ages and initial ¹⁷⁶Lu/¹⁷⁷Hf ratios for each zircon reveal radiogenic $\varepsilon Hf_{(t)}$ values between +4.5 to +7.4 (mean $\varepsilon Hf_{(t)} = 6.0 \pm 0.9$), corresponding to a relatively tightly grouped range of two-stage T_{DM}^{Hf} model ages between 2.25 and 2.45 Ga (Fig. 5.10a).

5.5.3.2 Foliated dacitic volcanoclasite (SB173B)

Sample SB173B zircon morphology is commonly euhedral and prismatic, and rarely pyramidal, with ratios of 1:2 to 1:4, 80–150 μ m in length and 50–75 μ m in width, with larger zircons reaching 200 μ m in length. Smaller prismatic grains are pale coloured and relatively free of fractures whilst larger grains show brown colouration and more extensive fracture networks. Cathodoluminescence images reveals fine, low-luminescent, oscillatory zoning in some grains with no clear evidence of xenocrystic cores, whilst others show broad zoning with low luminescence (Fig. 5.8b). BSE images show small mineral inclusions in a number of zircons and extensive fracturing in some smaller grains (likely to be the product of the mounting process).

Spot Name	204Pb /206Pb	±%	207Pb /206Pb	% ∓%	208Рb /206Рb	±%	206Pb /238U	%Ŧ	% 206Pb _c	U ppm	ppm Th	(1) ppm 206Pb*	232Th /238U	(1 /2 ±⁰% A) 208Pb 32Th <u>3</u> e	(1) Ag) 206Pb 38U ;e	(1) /20 Age	207Pb 6Pb		% Dis- cord- ant
SB023 Granite Accepted																					
S023-67	0.000065	45	0.1338	0.68	0.063	1.53	0.716	1.57	0.10	152	32	49	0.218	0.49	2024	49	2066	20	2137	13	+4
S023-69R	0.000065	29	0.1335	0.45	0.090	0.84	0.756	0.40	0.10	337	103	111	0.317	0.29	2056	29	2101	17	2133	9	+2
S023-70	0.000104	25	0.1337	0.49	0.120	0.80	0.709	2.08	0.15	309	126	100	0.422	0.28	1996	63	2055	55	2129	10	+4
S023-73R	I	ī	0.1326	0.54	0.077	1.09	0.709	1.02	0.00	258	71	87	0.284	0.35	2045	30	2124	18	2133	9	<u>+</u>
S023-76R	0.000168	32	0.1329	0.78	0.047	1.98	0.638	2.03	0.25	128	18	38	0.147	0.62	1883	90	1913	19	2107	17	+11
S023-77	0.000323	22	0.1353	0.76	0.071	1.61	0.677	1.07	0.48	134	30	4	0.235	0.52	1894	84	2089	21	2112	19	+
S023-79	0.000042	45	0.1322	0.56	0.153	3.77	0.641	2.43	0.06	257	133	81	0.535	0.28	1999	75	2010	17	2120	10	4
S023-81R	0.000155	35	0.1359	0.83	0.069	1.80	0.683	2.14	0.23	112	26	37	0.238	0.56	1961	70	2088	22	2150	17	$\overset{+}{\omega}$
S023-83R	0.000151	33	0.1357	0.77	0.053	1.90	0.640	1.39	0.22	132	22	38	0.169	0.58	1854	76	1888	19	2148	16	+14
S023-85	0.000143	19	0.1349	0.44	0.062	4.37	0.660	2.19	0.21	417	88	128	0.217	0.31	1812	83	1969	16	2138	9	+9
S023-86	0.000216	32	0.1363	0.88	0.064	1.96	0.673	2.33	0.32	103	22	35	0.221	0.62	1961	90	2155	23	2144	20	<u>-</u>
S023-87	0.000198	27	0.1338	0.72	0.068	1.53	0.638	1.85	0.29	158	33	46	0.214	0.48	1886	4	1887	18	2114	16	+12
S023-89R	0.000134	27	0.1345	0.59	0.045	1.54	0.603	2.42	0.20	250	45	73	0.185	0.43	1468	51	1899	17	2135	12	+13
S023-90	0.000114	28	0.1350	0.56	0.053	1.37	0.634	3.05	0.17	261	43	82	0.172	0.43	2001	54	2010	17	2145	Ξ	7+
S023-91	-0.000016	100	0.1337	0.76	0.025	15.48	0.645	2.45	-0.02	143	12	46	0.090	0.77	2057	307	2046	20	2149	14	+6
S023-93	0.000164	24	0.1376	0.59	0.114	1.00	0.619	0.49	0.24	231	87	68	0.391	0.32	1834	34	1893	16	2170	12	+15
S023-96	0.000082	71	0.1373	1.21	0.090	2.31	0.688	1.15	0.12	51	15	17	0.311	0.74	2116	75	2138	30	2180	23	+2
S023-98R	0.000197	25	0.1351	0.67	0.055	1.61	0.561	2.49	0.29	221	37	65	0.175	0.48	1823	71	1896	17	2131	15	+13
Cores																					
S023-69C	0.000258	38	0.1371	1.14	0.018	30.04	0.658	0.97	0.38	57	2	17	0.033	1.87	1871	851	1932	25	2147	26	+12
S023-73C	0.000226	45	0.1383	1.26	0.069	2.74	0.718	2.51	0.33	46	10	16	0.225	0.89	2067	130	2147	31	2168	28	<u>+</u>
S023-78C	-0.000040	100	0.1430	1.18	0.104	2.19	0.746	1.04	-0.06	51	17	18	0.340	0.72	2442	72	2241	31	2270	21	+2
S023-81C	0.000064	26	0.1365	0.39	0.102	0.71	0.729	2.20	0.09	487	174	164	0.369	0.25	2034	25	2137	16	2173	7	+2
S023-83C	-0.000282	41	0.1382	1.29	0.042	3.58	0.705	1.11	-0.41	43	5	14	0.131	1.13	2811	227	2072	30	2250	28	-9
S023-89C	0.000183	32	0.1331	0.82	0.063	1.84	0.645	0.69	0.27	127	27	42	0.221	0.56	1919	77	2117	22	2108	18	<u>-</u>
S023-98C	-0.000021	71	0.1337	0.64	0.014	2.96	0.654	0.60	-0.03	214	10	71	0.047	0.90	2230	103	2108	19	2150	11	+46
S023-99C	0.000121	38	0.1349	0.80	0.013	3.91	0.662	2.40	0.18	128	4	42	0.034	1.29	1831	338	2103	21	2142	16	+2
Errors are 1σ; Pb _c	and Pb* ind	licate the co	ommon and	radiogei	nic portions	, respecti	vely.		(1) Commo	on Pb co	rrected us in	g measured	l ²⁰⁴ Pb.							

Table 5.2. Isotopic data obtained by SHRIMP analyses at Curtin University, Perth, Australia

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Table 5.2. (Contin	(pən																				
														U) 208Pb	(1) 2	06Pb	(1)	07Pb	•`	6 Dis-
Spot Name	204Pb /206Pb	4%	207Pb /206Pb	7%∓	208Pb /206Pb	≁%	206Pb /238U	4%	% 206Pb _c	n U	ndq Th	(1) ppm 206Pb*	232Th /238U	 12 4;	32Th ge	/238 Age	n	/206 Age	Pb		cord- ant
SB023 - Rejected																					
S023-68			0.1356	0.99	0.022	3.74	0.588	4.21	0.00	86	9	22	0.069	1.07	1849	72	1702	20	2172	17	+25
S023-72	0.000266	18	0.1354	0.50	0.147	0.76	0.520	2.49	0.40	359	160	78	0.460	0.24	1479	24	1446	12	2124	12	+36
S023-76C	0.000969	71	0.1839	3.82	0.109	36.47	0.720	6.34	1.35	4	0	2	0.038	6.81	12394	4447	2223	98	2576	109	+11
S023-78R	0.000809	12	0.1435	0.54	0.093	1.02	0.481	2.33	1.20	350	95	75	0.281	0.30	1152	58	1435	12	2136	20	+37
S023-82	0.000368	22	0.1366	0.79	0.109	1.38	0.567	3.85	0.55	145	80	39	0.569	0.35	1031	34	1732	17	2122	20	+21
S023-92C	0.000285	41	0.1449	1.27	0.072	2.77	0.458	3.29	0.42	61	10	12	0.164	0.82	1656	115	1307	17	2242	29	
S023-92R	0.000154	33	0.1343	3.25	0.044	2.10	0.645	1.25	0.23	135	20	45	0.154	0.63	1875	94	2114	21	2128	59	-
S023-97	0.000147	35	0.1365	3.10	0.031	14.73	0.663	1.52	0.22	123	12	41	0.101	0.79	1953	303	2121	22	2159	56	+2
S023-98	0.000026	71	0.1332	0.69	0.053	1.67	0.481	1.93	0.04			0			ı		0	0	2136	12	
S023-99R	0.000489	18	0.1356	0.67	0.075	1.38	0.503	3.23	0.73	223	30	47	0.141	0.49	1984	104	1419	13	2086	20	+13
SB092 Quartz Die	orite																				
Accepted (all)																					
S092-01	0.000466	19	0.1406	0.91	0.160	1.15	0.752	1.9	0.69	105	55	36	0.55	0.47	2045	54	2178	25	2157	23	Ŀ
S092-02	0.000434	21	0.1398	0.81	0.192	1.11	0.749	1.5	0.64	96	61	33	0.66	0.46	2062	49	2177	26	2152	22	-1
S092-03	0.000256	21	0.1385	0.64	0.153	0.96	0.704	2.1	0.38	165	81	54	0.51	0.38	2085	39	2072	20	2166	15	+5
S092-04	0.000490	19	0.1374	0.81	0.166	1.18	0.797	2.2	0.73	94	52	32	0.57	0.48	2012	55	2170	26	2111	23	ų
S092-05	0.000205	24	0.1354	0.64	0.215	0.83	0.804	1.4	0.30	154	113	53	0.76	0.35	2096	32	2159	22	2134	14	-1
S092-06	0.000394	19	0.1401	0.70	0.216	0.92	0.768	2.4	0.58	131	94	44	0.74	0.39	2071	39	2134	23	2163	18	2 +
S092-07	0.000542	18	0.1398	0.71	0.237	06.0	0.780	1.9	0.81	127	101	43	0.82	0.39	2047	96	2160	93	2134	22	-1
S092-08	0.000428	19	0.1369	0.86	0.170	1.04	0.796	1.5	0.64	123	69	42	0.58	0.43	2047	47	2151	24	2115	21	-7
S092-09	0.000163	28	0.1380	0.66	0.149	1.01	0.795	1.9	0.24	158	62	53	0.52	0.40	2074	37	2128	22	2175	14	\tilde{c}^+
S092-10	0.000173	24	0.1365	0.58	0.142	06.0	0.816	1.4	0.26	199	76	99	0.50	1.83	2022	48	2109	20	2154	12	\dot{c}^+
S092-11	0.000126	26	0.1371	0.55	0.176	0.77	0.749	1.8	0.19	252	151	82	0.62	0.32	2012	68	2066	64	2170	11	9+
S092-12	0.000180	29	0.1377	0.73	0.187	1.01	0.745	2.2	0.27	140	88	45	0.65	0.40	2013	39	2064	27	2168	16	9+
S092-13	0.000170	20	0.1357	0.49	0.220	0.64	0.846	2.2	0.25	262	201	86	0.79	0.27	1981	24	2090	18	2144	11	\dot{c}^+
S092-14	0.000129	24	0.1366	0.50	0.230	0.64	0.806	2.5	0.19	266	212	85	0.82	0.27	1970	23	2049	18	2162	10	9+
S092-15	0.000599	18	0.1429	0.74	0.246	0.93	0.721	1.4	0.89	130	127	42	1.01	0.38	1635	36	2108	24	2165	23	9+
S092-16	0.000317	21	0.1386	0.70	0.214	0.91	0.850	2.2	0.47	134	76	43	0.74	0.39	1977	36	2056	23	2157	17	9+

-	`																				
												0		2 (1) 08Ph	3	yby)	Ξ	7Ph		Dis-
Spot Name	204Pb /206Pb	±%	207Pb /206Pb	±%	208Pb /206Pb	±%	206Pb /238U	±%	% 206Pb	D D	ppm Th	ppm 206Pb*	232Th /238U	/2 ±% A	132Th ge	/2: Ag	38U 3e	/20 Ag	6Pb e		cor- dant
SB173B - Daciti	c volcanoclas	ite																			
Accepted (all)																					
173B-01	0.000075	26	0.1357	0.42	0.142	0.65	0.739	2.6	0.11	340	167	109	0.51	0.25	1981	22	2026	16	2161	×	4
173B-02	0.000257	19	0.1387	0.56	0.084	1.10	0.743	2.7	0.38	190	50	63	0.27	0.41	2039	52	2079	20	2168	13	+4
173 B- 03	0.000086	29	0.1350	0.50	0.091	0.95	0.799	2.4	0.13	221	71	74	0.33	0.35	1987	32	2124	19	2149	10	+
173B-04	0.000145	26	0.1352	0.59	0.082	1.16	0.777	1.4	0.22	168	48	55	0.30	0.41	1912	42	2059	20	2142	12	+4
173B-05	0.000161	27	0.1352	0.64	0.247	0.78	0.804	1.6	0.24	138	117	46	0.87	0.34	2066	30	2109	23	2139	14	+
173 B- 06	0.000077	32	0.1354	0.53	0.109	0.91	0.794	2.8	0.11	209	78	68	0.39	0.34	2010	30	2057	19	2156	10	5+
173 B- 07	0.000130	24	0.1365	0.51	0.167	0.73	0.780	2.6	0.19	230	132	75	0.59	0.29	2003	26	2049	18	2161	10	5+
173B-08	0.000128	22	0.1365	0.46	0.150	0.71	0.711	3.5	0.19	293	151	94	0.53	0.27	1965	24	2022	17	2162	9	,
173B-09	0.000058	45	0.1374	0.76	0.081	1.29	0.781	1.9	0.09	137	38	47	0.29	0.46	2106	41	2146	23	2185	14	+2
173B-10	0.000062	27	0.1352	0.40	0.177	2.32	0.778	3.1	0.09	385	238	126	0.64	0.23	2010	48	2069	17	2156	8	+4
173 B- 11	0.000395	18	0.1380	0.67	0.115	1.20	0.804	2.2	0.59	128	49	43	0.39	0.44	1928	55	2113	24	2135	17	+
173 B- 12	0.000135	27	0.1362	0.58	0.127	0.95	0.780	2.0	0.20	177	75	59	0.44	0.37	2070	34	2093	21	2157	12	$^+3$
173B-13	0.000140	18	0.1371	0.40	0.144	0.62	0.772	1.2	0.21	383	185	128	0.50	0.24	2084	24	2105	16	2167	~	÷
Sample SB248 -	Trondhjemit	e																			
Accepted																					
S248-06	0.000083	30	0.1354	0.52	0.056	1.23	0.708	1.6	0.12	246	46	81	0.192	0.40	2039	4	2098	16	2155	10	+3
S248-09	0.000173	20	0.1374	0.49	0.093	0.91	0.712	1.6	0.26	275	85	90	0.318	0.31	1998	36	2073	16	2166	10	-5
S248-10	0.000102	24	0.1356	0.45	0.108	0.79	0.735	1.2	0.15	300	109	86	0.376	0.27	2032	28	2082	15	2155	9	+4
S248-14	0.000079	26	0.1345	0.44	0.070	0.92	0.719	1.3	0.12	337	80	110	0.245	0.31	2019	32	2079	15	2144	8	+4
S248-15	0.000122	30	0.1350	0.63	0.068	1.36	0.701	1.3	0.18	163	36	53	0.231	0.44	2037	51	2078	18	2143	13	+4
S248-16	0.000043	45	0.1355	0.55	0.061	1.25	0.721	1.3	0.06	208	43	68	0.215	0.40	2032	38	2084	17	2163	10	+4
S248-17	0.000141	20	0.1367	0.46	0.103	0.82	0.713	1.0	0.21	309	104	101	0.348	0.28	2069	31	2084	15	2162	10	+4
S248-20	0.000056	41	0.1346	0.58	0.069	1.24	0.663	1.2	0.08	221	52	74	0.242	0.39	2068	40	2130	18	2149	11	<u>+</u>
S248-21	0.000206	19	0.1358	0.52	0.119	0.88	0.656	2.0	0.31	263	103	84	0.403	1.25	1997	42	2046	17	2139	12	5+
S248-24	0.000026	71	0.1369	0.68	0.101	1.22	0.624	3.0	0.04	170	56	53	0.340	0.39	2048	35	1989	19	2184	12	+10
S248-25	-0.000010	100	0.1357	0.60	0.050	1.51	0.657	2.5 -		224	36	74	0.166	0.47	2221	42	2107	18	2175	11	+4
S248-27	0.000032	58	0.1348	0.62	0.082	1.23	0.689	1.4	0.05	193	54	64	0.288	0.40	2090	37	2106	19	2156	11	$^+_3$
S248-28	0.000072	30	0.1348	0.49	0.094	3.53	0.635	0.8	0.11	333	107	106	0.333	0.30	1970	70	2035	16	2149	6	6

Table 5.2. (Contin	(pənı																				
Sample SB248 - T	rondhjemite																				
Accepted (cont.)																					
S248-29	0.000257	20	0.1361	09.0	0.074	1.24	0.669	1.3	0.38	214	49	89	0.234	0.41	1981	09	2031	18	2134	14	9+
S248-32	0.000187	20	0.1363	0.52	0.092	0.98	0.685	2.2	0.28	280	83	87	0.306	0.33	1956	39	1988	17	2149	11	6+
S248-39	0.000096	25	0.1350	0.47	0.140	0.72	0.732	1.0	0.14	322	159	106	0.509	0.26	1983	26	2088	17	2147	6	1 3
S248-40	0.000138	23	0.1357	0.51	0.101	0.92	0.698	2.0	0.21	270	91	86	0.347	0.31	1989	34	2043	17	2149	11	9+
S248-48	0.000120	28	0.1353	0.58	0.062	1.30	0.647	3.3	0.18	231	49	73	0.220	0.40	1874	46	2020	17	2147	12	L+
S248-49	0.000039	45	0.1356	0.53	0.074	1.10	0.692	1.8	0.06	238	62	62	0.269	0.35	2007	32	2097	16	2165	10	+4
S248-51	0.000141	22	0.1363	0.51	0.063	1.13	0.635	2.3	0.21	309	74	16	0.248	0.33	1566	36	1893	16	2157	10	+14
Rejected																					
S248-05	0.000874	12	0.1508	0.58	0.117	6.10	0.717	2.1	1.28	184	47	61	0.265	0.40	2445	170	2118	18	2220	20	5+
S248-22C	0.000047	71	0.1371	4.27	0.062	2.13	0.674	1.8	0.07	90	17	30	0.197	0.68	2280	69	2131	24	2184	75	+3
S248-22R	0.000235	10	0.1228	0.28	0.061	09.0	0.425	4.5	0.36	2120	368	372	0.179	0.19	1192	21	1198	6	1951	7	+42
S248-38	0.000112	28	0.1367	2.19	0.085	1.09	0.590	1.9	0.17	269	90	71	0.345	0.31	1408	27	1728	15	2167	39	+23
S248-43	0.000098	21	0.1297	0.40	0.119	0.66	0.684	0.4	0.15	436	181	112	0.429	0.22	1576	19	1693	12	2077	~	+21
S248-55	0.000659	12	0.1436	2.13	0.094	4.82	0.742	1.6	0.98	202	49	68	0.252	0.38	2134	127	2118	17	2164	42	2+

	0.9	lσ						1.1	± 0.000026	Ισ							
	3.7	mean						3.9	0.281517	mean							
S092 - 15	3.2	4	23	2165	2.550014	2.39	1:1	3.7	0.281503	148112	1.481120	0.0220	77	0.000767	62	0.281535	SB092 - 15
S092 - 14	2.2	+6	10	2162	2.615175	2.43	1.2	2.6	0.281475	148111	1.481110	0.0310	108	0.001078	70	0.281519	SB092 - 14
S092 - 13	4.8	+3	Ξ	2144	2.451308	2.33	1.3	4.8	0.281547	148097	1.480970	0.0192	66	0.000659	74	0.281574	SB092 - 13
S092 - 12	3.2	6	16	2168	2.551125	2.39	0.9	3.8	0.281503	148110	1.481100	0.0215	73	0.000730	51	0.281533	SB092 - 12
S092 - 11	4.7	4	Ш	2170	2.454532	2.33	0.9	5.3	0.281545	148105	1.481050	0.0301	101	0.001012	50	0.281587	SB092 - 11
S092 - 10	3.4	+3	12	2154	2.540686	2.39	1.1	3.6	0.281507	148107	1.481070	0.0432	145	0.001450	63	0.281567	SB092 - 10
S092 - 9	2.9	$^{+3}$	14	2175	2.574122	2.40	1.0	3.6	0.281493	148133	1.481330	0.0277	88	0.000880	56	0.281529	SB092 - 9
S092 - 8	4.8	-2	21	2115	2.450803	2.33	0.9	4.1	0.281548	148152	1.481520	0.0186	60	0.000605	52	0.281572	SB092 - 8
S092 - 7	3.1	<u>'</u> _	22	2134	2.56012	2.40	0.8	2.8	0.281499	148132	1.481320	0.0249	78	0.000778	46	0.281531	SB092 - 7
S092 - 6	2.7	+2	18	2163	2.581811	2.41	1.3	3.2	0.281489	148136	1.481360	0.0249	82	0.000816	74	0.281523	SB092 - 6
S092 - 5	3.4	<u>-</u>	14	2134	2.536936	2.38	1.0	3.2	0.281510	148106	1.481060	0.0221	72	0.000724	55	0.281539	SB092 - 5
S092 - 4	2.7	ப்	23	2111	2.58322	2.41	0.9	2.0	0.281489	148187	1.481870	0.0098	32	0.000317	53	0.281502	SB092 - 4
S092 - 3	5.0	₽+ 2+	15	2166	2.436033	2.32	1.0	5.5	0.281553	148155	1.481550	0.0238	81	0.000813	57	0.281587	SB092 - 3
S092 - 2	5.0	<u>-</u>	22	2152	2.435042	2.32	1.2	5.2	0.281554	148172	1.481720	0.0173	53	0.000533	71	0.281576	SB092 - 2
S092 - 1	4.4	<u>-</u>	23	2157	2.47282	2.34	1.2	4.7	0.281537	148174	1.481740	0.0171	55	0.000548	71	0.281560	SB092 - 1
	2144 Ma																SB092
	15	0						0.3	0.000037	IO							
	3.8	nean		(S;	tertied core	ung mr	(excinc	4.I	0 000037	inean							
SU23-99U	3 4.5	+2	10	2142 م	2.40	2.34	0.1.0	4.0	0.281343	148044	1.480440	0.0531	1//	0.001770	50	0.281617	SB023 - 28
SU23-98C	0.8	+ ×	; =	0117	2.38	2.28	1.0	0.1	0.281581	148081	1.480810	0.0289	111	0.001114	2 28	0.281627	SB023 - 27
S023-96	- 33 0 - 80	5 +2	: 23	2180	2.51	2.38	1.2	4.7	0.281522	148086	1.480860	0.0977	321	0.003210	67	0.281655	SB023 - 25
S023-93	6.5	+15	12	2170	2.34	2.26	0.7	7.2	0.281600	148076	1.480760	0.0311	100	0.001000	39	0.281641	SB023 - 24
S023-91	5.0	+6	14	2149	2.43	2.32	0.9	5.3	0.281558	148071	1.480710	0.0319	101	0.001010	50	0.281599	SB023 - 21
S023-90	4.1	+7	Ξ	2145	2.49	2.35	0.6	4.3	0.281534	148049	1.480490	0.0442	147	0.001474	32	0.281594	SB023 - 20
S023-89C	4.5	Ļ	18	2108	2.46	2.33	1.1	3.9	0.281546	148079	1.480790	0.0290	87	0.000873	61	0.281581	SB023 - 19
S023-87	1.8	+12	16	2114	2.63	2.44	1.2	1.4	0.281471	148066	1.480660	0.0574	162	0.001624	68	0.281536	SB023 - 18
S023-86	4.5	<u>-</u>	20	2144	2.46	2.33	1.0	4.7	0.281544	148078	1.480780	0.0118	45	0.000454	56	0.281563	SB023 - 17
S023-85	3.9	+9	9	2138	2.50	2.36	0.8	4.0	0.281528	148053	1.480530	0.0377	130	0.001300	43	0.281581	SB023 - 16
S023-83R	4.3	+14	16	2148	2.47	2.34	1.1	4.6	0.281540	148100	1.481000	0.0221	69	0.000693	64	0.281568	SB023 - 15
S023-83C	4.1	$^{+9}$	28	2250	2.41	2.35	1.0	6.7	0.281533	148096	1.480960	0.0283	94	0.000940	55	0.281573	SB023 - 14
S023-81R	4.3	$^{+3}$	17	2150	2.47	2.34	0.8	4.6	0.281540	148104	1.481040	0.0165	57	0.000569	4	0.281563	SB023 - 12
S023-81C	1.8	+2	7	2173	2.63	2.46	0.8	2.6	0.281467	148090	1.480900	0.1088	339	0.003388	47	0.281607	SB023 - 11
S023-79	3.7	,	10	2120	2.52	2.37	0.9	3.3	0.281521	148103	1.481030	0.0203	67	0.000668	49	0.281548	SB023 - 10
S023-78C	1.2	+2	21	2270	2.59	2.47	1.0	4.1	0.281447	148084	1.480840	0.0465	144	0.001440	56	0.281509	SB023 - 9
S023-77	4.3	+	19	2112	2.48	2.34	0.8	3.7	0.281538	148097	1.480970	0.0216	82	0.000816	48	0.281571	SB023 - 8
S023-76R	3.2	+11	17	2107	2.55	2.38	0.8	2.6	0.281509	148074	1.480740	0.0276	100	0.001002	46	0.281549	SB023 - 7
S023-73C	2.4	+	28	2168	2.60	2.43	1.0	3.1	0.281484	148083	1.480830	0.0703	233	0.002326	58	0.281580	SB023 - 6
S023-70	3.8	+4	10	2129	2.51	2.36	1.0	3.6	0.281525	148100	1.481000	0.0396	140	0.001401	59	0.281582	SB023 - 4
S023-69C	2.0	+12	26	2147	2.62	2.44	0.9	2.2	0.281474	148103	1.481030	0.0614	194	0.001940	49	0.281553	SB023 - 3
S023-67	4.7	+4	13	2137	2.45	2.33	0.9	4.7	0.281549	148091	1.480910	0.0246	85	0.000850	52	0.281584	SB023 - 1
	2136 Ma																
spot	assigned age	(%)	±1σ	age (Ma)	(Ga)	(Ca)	±1σ	εHf(t)	¹⁷⁷ Hf (t)	(×10 ⁻⁵)	¹⁷⁷ Hf	¹⁷⁶ Yb/ ¹⁷⁷ Hf	(×10 ⁻⁵)	¹⁷⁷ Hf	(×10-6)	¹⁷⁷ Hf	Spot Name
SHRIMP	single	Disc.		²⁰⁷ Pb/ ²⁰⁶ Pb	crustal	T _{DM}			¹⁷⁶ Hf/	2SE	¹⁷⁸ Hf/		2SE	¹⁷⁶ Lu/	2SE	¹⁷⁶ Hf/	Sample
Corres.	sHf(t)	566 4 C 8 C			T _{DM}										0000275		1 40000
	at Curtin	alvsis	AS and	MC-ICP-A	hv $LA-\Lambda$	ned i	ohtai	iana (uthern Gh	area in so	the study	tic rocks in	тарта	data from	sotopic	Lu-Hfli	Table 5.3
_ Table 5.3:	continued																I
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Sample	¹⁷⁶ Hf/	2SE	¹⁷⁶ Lu/	2SE		¹⁷⁸ Hf/	2SE	¹⁷⁶ Hf/			T_{DM}		²⁰⁷ Pb/ ²⁰⁶ Pb		Disc.	single	SHRIMP
Spot Name	177 Hf	(×10-6)	¹⁷⁷ Hf	(×10 ⁻⁵)	¹⁷⁶ Yb/ ¹⁷⁷ Hf	¹⁷⁷ Hf	(×10 ⁻⁵)	177 Hf (t)	EHf(t)	±1σ	(Ga)		age (Ma)	±lσ	(%)	assigned age	spot
SB173b																2158 Ma	
SB173b - 1	0.281590	57	0.002180	218	0.0636	1.481150	148115	0.281500	3.5	1.0	2.40	2.55	2161	8	9+	3.4	173 B - 1
SB173b - 2	0.281522	57	0.001015	102	0.0332	1.480980	148098	0.281480	2.9	1.0	2.42	2.59	2168	13	+	2.7	173B - 2
SB173b - 3	0.281562	64	0.001974	197	0.0600	1.480820	148082	0.281481	2.5	1.1	2.43	2.59	2149	10	Ŧ	2.7	173B - 3
SB173b - 4	0.281626	68	0.001511	151	0.0485	1.481020	148102	0.281564	5.3	1.2	2.31	2.40	2142	12	+4	5.7	173B - 4
SB173b - 5	0.281579	67	0.002830	283	0.0891	1.481250	148125	0.281464	1.7	1.2	2.46	2.63	2139	14	+	2.1	173B - 5
SB173b - 6	0.281626	52	0.002500	250	0.0770	1.480680	148068	0.281523	4.2	0.9	2.37	2.50	2156	10	+5	4.3	173 B - 6
SB173b - 7	0.281579	52	0.000695	70	0.0228	1.481230	148123	0.281550	5.3	0.9	2.33	2.43	2161	10	+5	5.2	173B - 7
SB173b - 8	0.281540	61	0.000793	79	0.0232	1.481260	148126	0.281507	3.8	1.1	2.38	2.53	2162	6	9+	3.7	173B - 8
SB173b - 9	0.281568	68	0.000873	87	0.0238	1.481360	148136	0.281532	5.2	1.2	2.35	2.48	2185	14	+2	4.6	173B - 9
SB173b - 10	0.281548	73	0.001553	155	0.0472	1.481320	148132	0.281484	2.8	1.3	2.42	2.58	2156	8	+4	2.9	173 B - 10
SB173b - 11	0.281566	58	0.001230	123	0.0335	1.481110	148111	0.281516	3.5	1.0	2.38	2.51	2135	17	+1	4.0	173B - 11
SB173b - 12	0.281627	43	0.001460	146	0.0443	1.481010	148101	0.281567	5.8	0.8	2.31	2.40	2157	12	+3	5.8	173B - 12
SB173b - 13	0.281587	62	0.001250	125	0.0361	1.480690	148069	0.281535	4.9	1.1	2.35	2.47	2167	8	+3	4.7	173B - 13
							mean	0.281516	4.0						mean	4.0	
							lσ	0.000032	1.1						lσ	1.1	
SB248																2155 Ma	
SB248 - 1	0.281637	34	0.000531	53.1	0.0172	1.481040	148104	0.281615	7.4	0.6	2.24	2.29	2155	10	+3	7.4	S248 - 6
SB248 - 2	0.281630	67	0.000544	54.4	0.0179	1.480940	148094	0.281608	7.4	1.2	2.25	2.31	2166	10	+5	7.2	S248 - 9
SB248 - 3	0.281594	58	0.000389	38.9	0.0124	1.480710	148071	0.281578	6.1	1.0	2.29	2.37	2155	6	+4	6.1	S248 - 10
SB248 - 4	0.281648	42	0.000441	44.1	0.0142	1.480740	148074	0.281630	7.7	0.7	2.22	2.25	2144	×	+4	8.0	S248 - 14
SB248 - 5	0.281620	45	0.000519	51.9	0.0164	1.480620	148062	0.281599	6.6	0.8	2.26	2.33	2143	13	+4	6.9	S248 - 15
SB248 - 6	0.281591	53	0.000483	48.3	0.0154	1.480770	148077	0.281571	6.1	0.9	2.30	2.39	2163	10	$^{+4}$	5.9	S248 - 16
SB248 - 7	0.281584	48	0.000554	55.4	0.0179	1.480850	148085	0.281561	5.7	0.8	2.31	2.41	2162	10	+4	5.5	S248 - 17
SB248 - 8	0.281571	63	0.000505	50.5	0.0158	1.481050	148105	0.281550	5.0	1.1	2.32	2.44	2149	Π	Ŧ	5.1	S248 - 20
SB248 - 9	0.281570	99	0.000667	66.7	0.0216	1.480640	148064	0.281543	4.5	1.2	2.34	2.45	2139	12	+5	4.9	S248 - 21
SB248 - 10	0.281592	62	0.000580	58.0	0.0187	1.480650	148065	0.281568	6.4	1:1	2.30	2.40	2184	12	$^{+10}$	5.8	S248 - 24
SB248 - 11	0.281590	49	0.000412	41.2	0.0132	1.480660	148066	0.281573	6.4	0.9	2.29	2.38	2175	Π	+4	6.0	S248 - 25
SB248 - 12	0.281574	4	0.000276	27.6	0.0087	1.480720	148072	0.281563	5.6	0.8	2.31	2.41	2156	11	+3	5.6	S248 - 27
SB248 - 13	0.281598	42	0.000494	49.4	0.0152	1.480640	148064	0.281578	6.0	0.7	2.29	2.37	2149	6	9+	6.1	S248 - 28
SB248 - 14	0.281630	54	0.000725	72.5	0.0242	1.480580	148058	0.281601	6.4	0.9	2.26	2.32	2134	14	9+	6.9	S248 - 29
SB248 - 15	0.281601	50	0.000396	39.6	0.0130	1.480590	148059	0.281585	6.2	0.9	2.28	2.36	2149	Ξ	6+	6.4	S248 - 32
SB248 - 16	0.281575	55	0.000580	58.0	0.0183	1.480720	148072	0.281551	5.0	1.0	2.32	2.43	2147	6	\tilde{c}^+	5.2	S248 - 39
SB248 - 17	0.281583	45	0.000362	36.2	0.0109	1.480610	148061	0.281568	5.6	0.8	2.30	2.40	2149	Ξ	9+	5.8	S248 - 40
SB248 - 18	0.281585	37	0.000529	52.9	0.0163	1.480820	148082	0.281563	5.4	0.6	2.31	2.41	2147	12	L+	5.6	S248 - 48
SB248 - 19	0.281580	57	0.000533	53.3	0.0156	1.480480	148048	0.281558	5.6	1.0	2.31	2.42	2165	10	$^+$	5.4	S248 - 49
SB248 - 20	0.281569	46	0.000376	37.6	0.0116	1.480650	148065	0.281554	5.3	0.8	2.32	2.43	2157	10	+14	5.3	S248 - 51
							mean	0.281576	6.0						mean	6.0	
							lσ	0.000023	0.9						lσ	0.8	

Thirteen U-Pb analyses were performed on thirteen zircons with from both oscillatory zoned and unzoned zircons, which defined a single zircon age population. Analyses yielded a weighted average 207 Pb/ 206 Pb age of 2158 ± 6 Ma (MSWD = 1.02, n = 13) (Fig. 5.9c), interpreted as the crystallisation age of this rock, whilst a less precise upper intercept age of 2159 ± 14 Ma (MSWD = 1.11) was calculated from the concordia plot (Fig. 5.9d).



Fig. 5.9. Weighted average ${}^{207}Pb/{}^{206}Pb$ ages and Tera-Wassaberg concordia diagrams displaying SHRIMP U-Pb zircon spot data for samples SB248 (a,b), SB137B (c,d) and SB092 (e,f), displaying error bars of $\pm 1\sigma$. Accepted analyses shown in blue and rejected analyses shown in red.

Thirteen Hf analyses were performed on dated grains yielding ¹⁷⁶Lu/¹⁷⁷Hf of < 0.0028, ¹⁷⁶Yb/¹⁷⁷Hf of < 0.09 and ¹⁷⁶Hf/¹⁷⁷Hf range from 0.281464 to 0.281567 (± 0.000043 – 0.000073 2SE). A juvenile, radiogenic Hf character is indicated by a wider range of suprachondritic $\varepsilon Hf_{(t)}$ values between +1.7 and +5.8 (mean $\varepsilon Hf_{(t)} = 4.0 \pm 1.1$), corresponding to a range of T^{H f}_{DM} model ages spanning from 2.40 to 2.65 Ga (Fig. 5.10b).



Fig. 5.10. Results of combined U-Pb and Hf zircon analyses for samples from the study area in southwest Ghana, presented as ²⁰⁷Pb/²⁰⁶Pb age vs two-stage Hf model ages in billion years using the measure ¹⁶⁷Lu/¹⁷⁷Lu (first stage), a value of 0.0115 for the average continental crust (second stage) (Condie, et al., 2005) and a depleted mantle ¹⁷⁶Lu/¹⁷⁷Hf and ¹⁷⁶Hf/⁴⁷⁷Hf of 0.0384 and 0.28325 (Griffin, et al., 2002).

5.5.3.3 Hornblende quartz diorite (SB092)

Zircons for sample SB092 are subhedral to euhedral and clear to pale rose-coloured, with length to width ratio commonly between 1:2 and 1:3. Two morphologies are noted for SB092, one is more equant $(80-120 \ \mu m \log and 80-100 \ \mu m wide)$ and the other slightly more elongate and near prismatic (120-200 $\ \mu m \log and 50-80 \ \mu m wide)$. Inclusions of ilmenite needles are visible in some elongate grains. CL and BSE images show little evidence of metamictization. Most zircons have a large, uniform centre succeeded by irregular oscillatory zoning, a characteristic common to both morphological populations

(Fig. 5.8c). Sixteen zircon U-Pb analyses were performed on 16 zircons from both morphological groups. The two groups yielded indistinguishable ages, indicating a single zircon age population that yielded a weighted average 207 Pb/ 206 Pb age of 2155 ± 9 Ma (MSWD = 1.17, n =16) (Fig. 5.9e), which is interpreted as the crystallisation age. The Concordia plot yields an upper intercept age of 2144 ± 10 (MSWD = 0.81, n = 16) (Fig. 5.9f), however, a number of grains show reverse concordance, artificially lowering the intercept age.

Fifteen Hf analyses on fifteen grains yielded ¹⁷⁶Lu/¹⁷⁷Hf of < 0.003, ¹⁷⁶Yb/¹⁷⁷Hf of < 0.04 and ¹⁷⁶Hf/¹⁷⁷Hf range from 0.281475 to 0.281554 (± 0.000046 – 0.000074 2SE). These analyses yield a relatively radiogenic Hf signature, with suprachondritic ε *Hf*_(t) values ranging from +2.0 to +5.5 (mean ε *Hf*_(t) = 3.9 ± 1.1), which corresponds to a two-stage model age (T^{H f}_{DM}) range between 2.43 and 2.61 Ga, clustering in two groups at ~2.45 and ~2.56 Ga (Fig. 5.10c), suggesting two distinct mantle extraction events.

5.5.3.4 Biotite-hornblende potassic quartz monzonite (SB023)

Sample SB023 zircon morphology is commonly elongate, prismatic and clear to yellow colouring. Zircons are inclusion free and varied in size from 100 to 250 μ m along the *c*-axis. CL imaging revealed common xenocrystic inherited cores with both dull and bright luminescence, commonly with broad irregular zonation (Fig. 5.8d). Zonation in xenocrystic cores is often truncated along the rim with rare evidence of metamictization on core rims. Oscillatory zoning is common within thick (30 - 60 μ m) growth rings. Elongate grains without cores show pyramidal grain termination and regular oscillatory zoning. Large grains show more extensive cracking and fractures.

Thirty-six analyses were performed on twenty-six grains. Of the thirty-six analyses, four were performed on xenocrystic cores, two of which were rejected for common Pb above 1% and for ²³²Th/²³⁸U ratio <0.1. Two xenocrystic core analyses yielded ²⁰⁷Pb/²⁰⁶Pb ages of 2250 ± 28 Ma and 2270 ± 21 Ma. The remaining twenty-six analyses were performed on growth rims and homogeneous zircons. Eight of these analyses were rejected: one analysis had common Pb above 1%, two analyses yielded 1 σ > 50Ma and five grains had discordance > 15%. One population was identified from remaining 18 analyses yielding a weighted average ²⁰⁷Pb/²⁰⁶Pb age of 2136 ± 7 Ma (MSWD = 1.4, n = 18), which is interpreted to be the crystallization age of the rock (Fig. 5.10a). The Concordia diagram reveals an upper intercept age of 2138 ± 15 (MSWD = 0.82, n = 18) (Fig. 5.10b)), with the plotted from both concordant and

discordant accepted age indicating slight modern lead loss.



Fig. 5.11. Weighted average ${}^{207}Pb/{}^{206}Pb$ ages and Tera-Wassaberg concordia diagrams displaying SHRIMP U-Pb zircon spot data for sample SB023, displaying error bars of $\pm 1\sigma$. Accepted analyses shown in blue, rejected analyses shown in red, accepted analyses for inherited cores in green.

A total of twenty-two Hf analyses were performed on dated cores, rims and unzoned zircons from sample SB023, yielding ¹⁷⁶Lu/¹⁷⁷Hf of < 0.003, ¹⁷⁶Yb/¹⁷⁷Hf between 0.01 and 0.1, and ¹⁷⁶Hf/¹⁷⁷Hf range from 0.281447 to 0.281600 (± 0.000032 – 0.000068 2SE). Analyses revealed a varied, radiogenic Hf signature, with suprachondritic ε *Hf*_(t) values between +1.4 and +7.2 (mean ε *Hf* = 4.1 ± 0.6). The two inherited cores yielded strongly radiogenic Hf signatures, with values of ε *Hf*_(2250Ma) = +6.7 and ε *Hf*_(2270Ma) = +4.1, within range of the ε *Hf* values for homogeneous zircons and zircon rims in the same sample, for which there was no discernible systematic variation. The significant range of six epsilon units for all zircons analysed in sample SB023 is consistent with either heterogeneous magma sources or crustal contamination and corresponds to a wide range of T^H_{DM} values, from 2.34 to 2.63 Ga.

5.6 Discussion

5.6.1 Petrogenesis of magmatic suites

The observed major and trace element patterns of magmatic rocks are the product of source composition, petrogenetic mechanisms, magma mixing and fractionation crystallisation (e.g. Brown, *et al.*, 1984; Pearce, *et al.*, 1984a), as well as enrichment of magma sources and the residual phase assembly of partial melting, indicative of the pressure of melt production (e.g. Defant & Kepezhinskas,

2001; Foley, *et al.*, 2002; Moyen, 2011; Moyen & Stevens, 2006a; Xiong, *et al.*, 2005). The ubiquitous negative Nb-Ta, P and Ti anomalies coupled with positive Pb anomalies and variable LILE enrichment for all samples analyses in this study, excluding one meta-gabbro and one metabasalt, are consistent with the trace element patterns of the present-day bulk continental crust (Hawkesworth & Kemp, 2006a; Hofmann, 1997; Rudnick & Gao, 2003). Such features typify arc magmas and crust formation in a subduction setting (e.g. Arculus, *et al.*, 1999), and are therefore consistent with the onset of "modern-style" subduction-related processes after 3.0 Ga (e.g. Condie & Benn, 2006; Halla, *et al.*, 2009; Laurent, *et al.*, 2014; Smithies & Champion, 2000). Magmatic rocks sampled in this study share a number of geochemical similarities with late-Archaean granitoids generated within nascent subduction-related geodynamic settings, which are characterised by: (i) sodic tonalite-trondhjemite-granodiorite (TTG) granitoids derived from partial melting of hydrous metabasalts; (ii) mantle-derived, incompatible element-rich sanukitoids; (iii) biotite and two-mica granites derived from crustal sources; and, (iv) hybrid granitoids sharing petrological and geochemical similarities and sources to the first three groups (Halla, *et al.*, 2009; Laurent, *et al.*, 2014; Moyen, 2011; Moyen & Martin, 2012), suggesting somewhat comparable petrogenetic processes.

The two TTG groups, defined in section 5.1, are both characterised by a silicic, plagioclaserich assemblage, distinguished from one another on the basis on the dominance of either biotite or hornblende as the primary ferromagnesian mineral. They show tholeiitic to calc-alkaline affinities and low K_2O/Na_2O ratios, with higher ratios in the hornblende-bearing TTGs reflecting slightly higher modal proportions of potassium feldspar. Based on experimental melt compositions (Laurent, *et al.*, 2014, and references therein), Figure 5.12a indicates a common low-K mafic source for both TTG groups, with the exception of two samples, with one indicating a high-K mafic source, and the second high- Al_2O_3 sample indicating remelting of pre-existing TTGs. Although both groups of TTGs have low potassium values, normalised LREE/HREE ratios (Fig. 5.12b) indicate the hornblende-bearing granitoids show similar ratios to modern day basalt-andesite-dacite-rhyolite (BADR) arc sequences, where magma is primarily derived through metasomatism of mantle peridotite by fluids released during dehydration of the subducting slab (Martin, *et al.*, 2009). The complementary trace element patterns of the two groups of TTGs suggests a link in their magmatic history, but contrasting petrogenetic processes.



Fig. 5.12. a) Ternary diagram for major element of granitoid samples analysed in this study used to discriminate between potential source rocks, with fields after Laurent, et al. (2014, and references therein) based on experimental melt compositions. Binary diagrams (b-f) used to discriminate: (b) interaction and role of fluids or enriched TTG magmas in mantle-derived magmas (fields based onMartin, et al., 2009, and references therein); (c, d, e) depth of melt generation and residual phase assemblage, comparing late Archaean TTGs and post-2.5Ga TTG and arc magmatism (Drummond

& Defant, 1990; Moyen, 2011); and (f) granite discrimination plot for tectonic setting of magma generation (Pearce, et al., 1984a).

REE patterns of the biotite-dominant TTG group display a strongly fractionated REE pattern, with a significant depletion in HREE indicating residual garnet and/or hornblende in the magma source.

High Sr/Y ratios and Y content also reflect a garnet-bearing residuum (Fig. 5.12c), often characterising deep melting at a convergent margin (Condie, 2008; Martin, 1993). The La_N/Yb_N vs Yb_N binary diagram (Fig. 5.12d) indicate high modal proportions of garnet in residuum of the garneteclogite source for the biotite-rich TTGs, compared to the garnet-free amphibolite indicated for the hornblende-dominant TTGs (Drummond & Defant, 1990; Laurent, et al., 2014; Moyen, 2011). These high-pressure TTG melts display positive Sr and Eu anomalies (Fig. 5.12e), a characteristic commonly documented in slab-derived tonalite and trondhjemite melts involving subduction of young, hot oceanic crust due to the release of Sr and Eu during plagioclase breakdown at the garnet amphiboliteeclogite transition (Defant & Drummond, 1990; Drummond, et al., 1996; Drummond & Defant, 1990), occurring at pressures between 1.5 and 2.0 GPa (Moyen & Stevens, 2006b; Rapp, et al., 1991). As such, these magmas are considered analogous to adakites generated through the partial melting of a hot, down-going slab in an oceanic environment (Defant & Drummond, 1990; Drummond, et al., 1996; Smithies, 2000). Their unusually low Mg# and high SiO₂ and elevated concentrations of Sc, V and Co suggest that evidence of interaction and assimilation with peridotite in the overlying mantle wedge has been partially obscured by subsequent amphibole fractionation (Rapp, et al., 1999; Wang, et al., 2006; Woodhead, et al., 1993).

Predominantly low-K mafic sources are indicated for hornblende-bearing TTGs (Fig. 5.12a), however, compared to the biotite-bearing TTGs, they display lower Al_2O_3 contents, low Sr concentrations, weakly fractionated REE patterns with high HREE concentrations and slightly negative Eu anomalies. These characteristics indicate that magma production occurred within the stability field of plagioclase, indicating either plagioclase fractionation or plagioclase as a residual phase in a garnet-free or garnet poor-residuum at pressures of <1.0 GPa (Moyen & Stevens, 2006b). These characteristics suggest a shallow, low-pressure source, such as partial melting of garnet-free basaltic crust, interpreted as analogous to the low-pressure TTG group of Moyen (2011).

Contrasting TTG magmas, both the pyroxenite-gabbro and the diorite suites follow the calc-alkaline differentiation trend in the Na-K-Ca ternary diagram (Fig. 5.5). They display low SiO_2 contents, high FeO₁, MgO and Mg#, and enrichment in transitional metals including Cr and Ni, thus precluding melt

derivation from a purely crustal source. This is further highlighted by REE patterns which display less than an order of magnitude enrichment in HREE and MREE relative to primitive mantle values. Trace element patterns do, however, indicate moderate LILE and Pb enrichments for the quartz diorite suite, suggesting a crustal influence (Brown, *et al.*, 1984; Hawkesworth & Kemp, 2006a). The pyroxenite suite potentially represents the primary magma from which the dioritic magmas evolved through fractional crystallisation during differentiation, generated thorough partial melting of an enriched mantle source. Depletion of HFSE elements, including Nb, Ta, Zr and Hf, in diorite and quartz diorite samples is typical of island arc magmas generated during dehydration melting with a rutile-bearing residuum (Foley, *et al.*, 2002). Despite similarly low SiO₂, elevated ferromagnesian (FeO_T + MgO) content and Sr and Ba concentrations > 500 ppm, the diorite and quartz diorite samples are not considered to be analogous to the sanukitoids series frequently recorded at the Archaean-Palaeoproterozoic boundary, due to lower MgO concentrations, Mg# and weaker REE fractionation (Martin, *et al.*, 2009; Martin, *et al.*, 2005; Shirey & Hanson, 1984; Stern, *et al.*, 1989).

Three samples plot in the high-K calc-alkaline field, including the biotite-hornblende quartz monzonite and two granites. The quartz monzonite sample displays a similar composition to the highpressure TTGs in terms of REE distribution and a positive Sr anomaly, but features a higher K₂O/ Na2O ratio, greater concentrations of Rb, Ba and K, as well as lower SiO2 and no Eu anomaly. It may represent the mixture of magmas from a metasomatised mantle source and partial melting of preexisting high-pressure TTG sources (Feng & Kerrich, 1992). It contrasts the two-mica granite and the muscovite-bearing leucogranite samples, which display higher SiO₂ contents with elevated K₂O/ Na₂O (0.93–1.01), reflecting higher modal proportion of potassium feldspar. These granites display strongly fractionated REE patterns, substantial enrichment in LILE and negative Eu anomalies (Fig. 5.12e), indicating plagioclase fractionation or melt generation within the stability field of plagioclase, consistent with geochemical characteristics of S-type granites generated through partial melting of metasedimentary rocks (Clemens, 2003; Clemens & Vielzeuf, 1987). These granites are interpreted as the product of shallow partial melting of enriched crustal sources, including sedimentary packages, further supported by field observations (Fig. 5.3g), and calculated metamorphic conditions for anatectic terranes detailed in Chapter 4. Within the Sefwi Belt, the extremely low REE content of a sampled rhyolite (SB076) is a reflection of its petrological assemblage which contains no ferromagnesian phases, however, field relationships indicate that this rhyolite non-conformably overlies a garnet-amphibolebearing paragneiss and potentially represents a felsic melt due to partial melting of a nearby crustal

source. Figure 5.12f shows that the two-mica and muscovite bearing granites plot in the collisional granite field (Pearce, *et al.*, 1984a), whilst both TTG groups and the diorite groups plot in the volcanic arc granite field due to lower LILE enrichment.

5.6.2 Timing of juvenile crust formation, recycling and basin deposition

The temporal evolution of magmatism in the study indicates partly contemporaneous emplacement of high-pressure TTGs derived from melting of the subducting, young, hot, oceanic lithosphere, and quartz diorite melts derived from an enriched mantle source. These granitoids are emplaced into volcanic rocks of the Sefwi Belt, which have crystallisation ages between ca. 2189 and 2164 Ma (Hirdes & Davis, 1998; Hirdes, *et al.*, 2007). The dacitic volcaniclastic sample from the Sunyani Basin yielded a zircon U-Pb crystallisation age of 2158 \pm 6 Ma and displays similar trace element patterns to the lowpressure TTGs, and is therefore interpreted as their volcanic equivalent. Hornblende-bearing tonalites emplaced in the Ivorian extent of the Sefwi-Wiawso domain at 2162 \pm 1 Ma display moderate LILE enrichment relative to LREE and weak to intermediate REE fractionation patterns (Hirdes, *et al.*, 2007), and are also considered analogous to the hornblende-bearing TTG suite of this study. The timing of emplacement of the pyroxenite and gabbro suite remains poorly understood however, samples documented in this study are consistent with pyroxenites documented in south-eastern Ivory Coast also indicating a LILE-enriched upper mantle source (Hirdes, *et al.*, 2007).

Approximately 20 to 25 Myrs after TTG and diorite emplacement, high-K quartz monzonites yield zircon crystallisation ages of 2136 ± 7 Ma, interpreted as the product of remelting of existing high-pressure TTGs by mantle derived magmas. Quartz monzonite magmatism in the study area coincides with the emplacement of monzonitic and arfvedsonite-bearing albite granite magmatism between ca. 2145 and 2135 Ma in the Ivorian extent of the Sefwi-Wiawso Domain (Hirdes, *et al.*, 2007).

Finally, two-mica and muscovite granitoids are emplaced in the Sunyani-Comoé Domain along the north-western margin between ca. 2092 and 2081 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007; Petersson, *et al.*, 2016), associated with crustal thickening and anatexis of felsic or supracrustal rocks, as detailed in Chapters 3 and 4. The muscovite granite sample (SB010) is taken from the margin of a muscovite-bearing granitoid that yielded Pb-Pb ages of 2088 \pm 1 Ma (Hirdes, *et al.*, 1992) and the two-mica granite (SB370) was sampled from the Wenchi pluton dated at 2092 \pm 2 Ma (Agyei Duodu, *et al.*, 2009), suggesting that true granite magmas occur significantly later in the magmatic history of the study area.

The evolution of magmatic suites in the Neoarchaean terranes defines the following progression: (1) TTG magmatism, generated through the partial melting of mafic rocks; (2) crust-derived biotite or two-mica granitic magmatism; (3) sanukitoids derived from a hybridized incompatible element-rich source and mantle peridotite, and (4) high-K hybrid granites displaying characteristics of the three previous suites (Feng & Kerrich, 1992; Laurent, *et al.*, 2014; Smithies & Champion, 2000). Laurent, *et al.* (2014) contend that the distinct evolution and diversity of magmatism in Archaean granitoids represents an ancient manifestation of subduction and tectonic collision. Despite the absence of sanukitoid-like rocks in the study area, a similar diversity and evolution is hypothesised for southern Ghana, involving early subduction-related TTG magmatism from both the subducting slab and shallow partial melting of a basaltic source with coeval dioritic magmatism derived from the metasomatised mantle wedge, emplacement of minor hybrid, high-K plutons, followed by production of crust-derived peraluminous granites during collision of continental fragments in the early Palaeoproterozoic.

The results of U-Pb dating of magmatic zircons constrain a period of magmatic activity in the Sefwi Greenstone Belt and the Sunyani-Comoé Domain yield two inherited zircon ages of 2250 ± 28 and 2270 ± 21 Ma and crystallisation ages between ca. 2158 and 2136 Ma. Previous studies record a wider range of U-Pb crystallisation ages, between ca. 2222 and 2081 Ma, both within the study area and in the immediately adjacent Palaeoproterozoic domains (e.g. Davis, *et al.*, 1994; Delor, *et al.*, 1992; Hirdes & Davis, 1998; Hirdes, *et al.*, 2007; Oberthür, *et al.*, 1998; Siméon, *et al.*, 1992). The oldest preserved crystallisation age in the study area, yielded by a meta-gabbro in the Hwidiem Domain, for which K-Ar dating of amphibole indicating cooling below 550°C at 2222 ± 32 Ma (Feybesse, *et al.*, 2006), within error of inherited zircon cores dated in this study. Andesitic and rhyolitic flows and volcaniclastic rocks within the Sefwi Belt yield U-Pb and Pb-Pb zircon ages between ca. 2189 and 2164 (Hirdes & Davis, 1998; Hirdes, *et al.*, 2007), partly contemporaneous with the emplacement of "belt-type" or "Dixcove-type" TTG granitoids into volcanic rocks of the Sefwi Belt between ca. 2179 and 2154 Ma (Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007; Petersson, *et al.*, 2017).

On the northern margin of the Sefwi Greenstone Belt, the timing of intermediate volcanism and sedimentation in the Sunyani-Comoé Domain (also referred to in the literature as the Sunyani Basin, or an extension of the Comoé Basin; Agyei Duodu, *et al.*, 2009; Feybesse, *et al.*, 2006; Vidal & Alric, 1994) is constrained by syn-depositional dacitic sample SB173B, which yielded an 207 Pb/ 206 Pb crystallisation age of 2158 ± 6 Ma. This dacite is found intercalated with low-grade volcano-sedimentary rocks and provides a new maximum age for the commencement of sedimentation for the domain, previously constrained by post-depositional emplacement of elongate leucogranites between ca. 2092 and 2081 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007). Furthermore, this age indicates that volcanism and sedimentation in the Sunyani-Comoé Domain was contemporaneous with plutonism in the Sefwi Belt, suggesting coeval crustal formation with an unknown initial spatial relationship. Comparison of the timing of deposition of volcano-sediments in the Kumasi-Afema Domain, bracketed between ca. 2135 and 2122 Ma (Adadey, *et al.*, 2009) or between ca. 2154 to 2136 Ma (Perrouty, *et al.*, 2012) with the 2158 \pm 6 Ma syn-depositional dacitic volcanism from the Sunyani-Comoé Domain (this study) indicates diachronous basin development with sediment deposition commencing between 10 and 30 Myrs earlier in the Sunyani-Comoé Domain. This finding would support the timing of extension event between ca. 2145 and 2135 Ma, hypothesised by Hirdes, *et al.* (2007), potentially responsible for the original basin opening of the Kumasi-Afema Domain.

The original spatial relationships between the Sefwi Belt and the Sunyani-Comoé Domain is largely obscured by deformation during the Eburnean Orogeny, however, we proposed that the tectonic contacts between the two domains represents a narrow collisional front or suture, with evidence of collision between the two discrete crustal fragments, represented by southern Ghana and central Ghana/Ivory Coast, recorded by the emplacement of syn-collisional granites (Fig. 5.12f) between ca. 2092 and 2081 Ma, and high-P, intermediate-T metamorphism and clockwise P-T metamorphic paths (Ch. 4).

Zircons from magmatic rocks analysed in this study display suprachondritic ϵ Hf values between +1.2 and +8.0 at the time of zircon crystallisation reflecting a highly radiogenic source and the juvenile nature of the newly formed crust. Furthermore, the juvenile Hf signature of all samples suggests that older, Archaean crustal material was absent or distal during juvenile crust production, consistent with the findings of previous studies of Palaeoproterozoic rocks in the West Africa Craton that yielded positive ϵ Nd between +1.2 and +4.3 (Abouchami, *et al.*, 1990; Boher, *et al.*, 1992; Gasquet, *et al.*, 2003; Taylor, *et al.*, 1992). The range of both ϵ Hf values and $T_{\rm DM}^{\rm Hf}$ model ages suggest slightly heterogeneous magma sources for magmatism at ca. 2155 Ma, with greater ranges of ϵ Hf values and $T_{\rm DM}^{\rm Hf}$ model ages in the ca. 2136 Ma quartz monzonite suggesting increasingly heterogeneous sources late in the magmatic evolution, in agreement with geochemical analysis. Two-stage Hf model ages for all samples indicate the juvenile crust was extracted from the mantle between ca. 2260 Ma and 2650 Ma (Fig. 5.13), indicating a Neoarchaean maximal age of a reworked proto-crust of 2650 Ma, from which the Palaeoproterozoic crust was derived. When considering both inherited zircon cores and crystallisation ages of magmatic rocks in the study area, crustal residence times vary significantly between 100 and 400 million years.



Fig. 5.13. ε Hf values versus Pb-Pb age diagram for individual zircon grains from the study area. Average continental crust evolution is modelled with ¹⁷⁶Lu/¹⁷⁷Hf = 0.0115 (Condie, et al., 2005) and a mafic crust evolution with ¹⁷⁶Lu/¹⁷⁷Hf = 0.0210 (Kramers, 2007). The two evolutionary trends proposed by different authors (Blichert-Toft & Puchtel, 2010; Griffin, et al., 2002) for the composition of the depleted mantle display similar results for Hafnium model ages, represented by stars at the intersection of depleted mantle and crustal evolution trends.

Two conclusions may be drawn from the combined geochemical and isotopic characteristics of igneous rocks in the Sefwi Belt. Firstly, the high-pressure TTG group, represented by ca. 2155 Ma trondhjemite sample (SB248), was generated through the partial melting of a subducting slab, extracted from the depleted mantle between approximately ca. 2300 and 2260 Ma. Secondly, diorites, quartz diorites and younger quartz monzonites have a wider range of ϵ Hf values, consistent with continued contribution of radiogenic material from the depleted mantle at ~ 2250 Ma combined with recycled juvenile proto-crust with mantle extraction ages spanning from ~2650 to 2300 Ma. Within the Baoulé-Mossi domain of the WAC, inherited zircon ages span from the Mesoarchaean to Palaeoproterozoic, with approximately 10 grains in southern Mali presenting inherited ages between ca. 3600 and 2700 Ma proximal to the Archaean Kénéma-Man Domain (Parra-Avila, *et al.*, 2016). Younger inheritance is recorded by the Ghanaian Gondo granite gneiss, dated at ca. 2876 Ma, (Thomas, *et al.*, 2009) and the Infantayire granite gneiss with inherited ages between ca. 2390 and 2260 Ma (Siegfried, *et al.*, 2009), and the 2312 \pm 12 Ma inherited zircons age from the Dabakala tonalite, Ivory Coast (Gasquet, *et al.*, 2003). By comparison, the oldest recorded crystallisation ages within the Baoulé-Mossi Domain is 2265 \pm 17 Ma, yielded by a granodiorite in the Goren/Po-Tenkodogo-Yamba of the Banfora Belt,

Burkina Faso (Parra-Avila, *et al.*, 2017), and 2253 \pm 9 Ma, yielded by a granodiorite gneiss in northeast Burkina Faso (Tshibubudze, *et al.*, 2013). With the exception of the Gondo granite gneiss, these ages are comparable to inherited ages yielded by the quartz monzonite in this study. The lack of rocks with ages between ca. 2650 and 2300 Ma preserved in the Baoulé-Mossi Domain suggests that the juvenile proto-crust from which the domain was derived was completely recycled prior to the oldest crystallisation age in the Sefwi Belt of 2222 \pm 32 Ma (Feybesse, *et al.*, 2006). Alternatively, crust older than ca. 2222 Ma may be as yet undocumented or extensively eroded, as suggested by the presence of older zircons in stream sediments in southern Mali.

5.6.3 Implications for the crustal architecture of the West African Craton

A database compilation of Hf-isotope data for igneous and detrital zircons sourced from Palaeoproterozoic rocks within the West African Craton is illustrated in Figure 5.14. Figure 5.14 contains all $\epsilon Hf_{(i)}$ values currently available for WAC (pale grey diamonds) (data sourced from: Block, *et al.*, 2016a; Eglinger, *et al.*, 2017; Parra-Avila, *et al.*, 2016; Petersson, *et al.*, 2017; Petersson, *et al.*, 2016), demonstrates the spectra of values within and between discrete samples, highlights a 150 M.y. period of juvenile crust production between \sim 2250 and 2100 Ma, following a period of reworking of Archaean crust for zircons aged between ca. 3200 and 2600 Ma. $\epsilon Hf_{(i)}$ values for the Sefwi Belt (this study; Fig. 5.14) are consistently radiogenic, spanning from with +1 to +8, consistent with published ranges from the Palaeoproterozic domain of the West African Craton.



Fig. 5.14. Compilation of all $\varepsilon Hf_{(i)}$ values calculated for igneous and detrital zircons from the Baoulé-Mossi domain of the West African Craton (pale grey diamonds), relative to the evolution of the depleted mantle composition of (Griffin, et al., 2002), and values yielded by this study for igneous samples SB023, SB092, SB173B and SB248. Data sourced from Block,

The most significant implication of this data is realised when it is displayed as an overlay of the geological map of the Baoulé-Mossi Domain (Fig. 5.15), highlighting the spatial distribution of juvenile versus reworked crustal material. Whilst we acknowledge the relative sparsity of the data, it may be noted that the highly radiogenic Hf signatures of ca. 2155 Ma samples from the Sefwi Belt (this study; Petersson, *et al.*, 2017) separate the sub-chondritic Hf signature of the Kibi Belt from the sub-chondritic to supra-chondritic rocks of the Maluwe and Sawla Suite of NW Ghana (Petersson, *et al.*, 2017), and the more distal mixed Hf signatures near the Archaean-Palaeoproterozoic boundary in southern Mali (Parra-Avila, *et al.*, 2016) and in Palaeoproterozoic rocks in Guinea external to the Archaean domain (Eglinger, *et al.*, 2017).



Fig. 5.15. $\varepsilon Hf_{(i)}$ values yielded by Palaeoproterozoic rocks of the West African Craton (Data sources: Block, et al., 2016a; Eglinger, et al., 2017; Parra-Avila, et al., 2016; Petersson, et al., 2017; Petersson, et al., 2016) highlighting the isotopic boundary (blue rectangle), represented by the Sefwi Belt in southwest Ghana, and significant reworked Archaean components in southeast Ghana, southern Mali and Guinea. Simplified greyscale map of the southern portion of the West African Craton from WAXI3.

We, therefore, propose that the juvenile nature of crust exposed in the study area represents an isotopic boundary within the crustal architecture of the WAC. Testing of this hypothesis, however, requires significantly more data, especially in the Ivory Coast. This data supports the conclusions of a number of authors who suggest that isotopic signatures of magmatic rocks in southern Ghana require a second, as yet unknown, Archaean crustal source that lay southeast of the Baoulé-Mossi domain during crust formation (Petersson, *et al.*, 2017; Taylor, *et al.*, 1992). Further isotopic mapping of internal boundaries within the juvenile terranes of the WAC is somewhat limited by the capability of Hf analyses to recognise the reworking of recently created crust (Payne, *et al.*, 2016, and references therein), thus impeding our understanding of the crustal architecture through Hf-isotopes alone.

5.6.4 A geodynamic model for the southern West African Craton

Palaeoproterozoic greenstone belts and granite-gneiss domains in the West African Craton have been likened to Archaean provinces based on lithological and structural associations (e.g. Delor, *et al.*, 1995; Lompo, 2009; Lompo, 2010; Vidal, *et al.*, 1996; Vidal, *et al.*, 2009), whilst others consider greenstone belts in West Africa as analogous to modern volcanic arcs (e.g. Baratoux, *et al.*, 2011; Dampare, *et al.*, 2008; Senyah, *et al.*, 2016; Sylvester & Attoh, 1992). New geochemical and isotopic data presented in this study supports subduction-related magma generation, with most of the magmatic suites analysed from the Sefwi Greenstone Belt bearing geochemical signatures characteristic of arc environments. The proposed arc type environment is consistent with a growing body of evidence for volcanic arcs generated in a subduction setting within the West African Craton (Ama Salah, *et al.*, 1996; Béziat, *et al.*, 2000; Block, *et al.*, 2016a; Lambert-Smith, *et al.*, 2016; Parra-Avila, *et al.*, 2017; Petersson, *et al.*, 2016; Sylvester & Attoh, 1992; Tapsoba, *et al.*, 2013).

The two end-member models generally considered for the production of TTGs through partial melting of hydrous metabasalts are distinguished by two contrasting geodynamic settings. The first geodynamic setting for TTG production is through partial melting of a basalt underplating thickened crust (Albarède, 1998; Atherton & Petford, 1993; Bédard, 2006; de Wit, 1998; Smithies, *et al.*, 2005). The second setting is through the partial melting of a young, hot subducting slab, rather than dehydration melting (Foley, *et al.*, 2002; Martin, 1986; Martin & Moyen, 2002; Rapp & Watson, 1995; Rapp, *et al.*, 1991), which generates melts that more closely resemble the geochemical characteristics of the high-P, low-HREE TTGs occurring in the Sefwi Belt.

Juvenile magmatism documented in the study area is likely sourced from both the metasomatised mantle wedge as well as the subducting slab. This suggests that the geodynamic setting for crust production was hot enough to generate slab-derived melts, more than 250 Myrs younger than the 2500 Ma end-date of slab derived melts, proposed by Martin, *et al.* (2009), reflecting a decrease in the Earth's heat production. We note the diversity and chronological evolution of magmatism in the study area, including subduction-related magmatism followed by emplacement of syn-collisional crustal melts and hybrid granitoids, which somewhat resembles those recorded in juvenile terranes from the late Archaean, including the Pietersburg Block of the Kaapvaal Craton (Laurent, *et al.*, 2014) and the Kola and Karelian Craton (Halla, *et al.*, 2009), elucidated by similar petrogenetic processes and broadly comparable geodynamic settings. In contrast to these systems however, the crustal evolution documented here in southern Ghana occurs in <100 Myrs, as opposed to the 1000 Myrs evolutionary period proposed for Archaean terranes (Laurent & Zeh, 2015).

Whilst magmatism in the Sefwi Belt and the adjacent domain is consistent with the peak magmatic activity for the eastern block of the West African Craton (Parra-Avila, et al., 2017), there are multiple avenues of evidence which suggest southern Ghana docked onto central Ghana and the central east Ivory Coast at ca. 2090 Ma. Evidence includes collision-related crust-derived granitic magmatism along the north west margin of the Sefwi Belt between ca. 2092 and 2081 Ma (this study; Hirdes, et al., 1996; Hirdes, et al., 2007) as well as high-pressure, intermediate-temperature metamorphism and anatexis (Ch. 4) with retrograde (post-peak T) metamorphism constrained by in-situ monazite U-Pb ages dated at ca. 2073 Ma. Within the broader context of the Palaeoproterozoic Baoulé-Mossi Domain of the WAC, diachroneity in the geochronological record suggests additional crustal boundaries coincide the boundary located between the Banfora and Bagoe belts of western Burkina Faso and southern Mali, respectively, as identified by Parra-Avila, et al. (2017). Block, et al. (2016b) recently proposed a collisional event at ca. 2130 Ma between NW Ghana and western Burkina Faso/Ivory Coast associated with peak metamorphism and crustal anatexis. This coincides with peak magmatism and the first phase of the westward migration of magmatism from the eastern most portions of the Baoulé-Mossi domain to the central crustal boundary (Hirdes, et al., 1996; Parra-Avila, et al., 2017). Parra-Avila, et al. (2017) note a second westward migration of magmatism at ca. 2090 Ma, with subsequent synchronous subduction-related juvenile crust formation of the Kédougou-Kéniéma Inlier (Hirdes & Davis, 2002a; Lambert-Smith, et al., 2016) and high-K magmatism along the northern margin of the Kénéma Man Shield (Egal, et al., 2002; Eglinger, et al., 2017) during the collision of the Archaean and Palaeoproterozoic domains. This protracted westward migration of magmatism and subsequent cessation of magmatism is incompatible with the timing of collision between the Kibi Belt and an Archaean fragment at ca. 2140 Ma (Petersson, et al., 2017), and the subsequent collision along the NW margin of the Sefwi Belt between southern Ghana and central Ghana/eastern Ivory Coast between 2100 and 2080 Ma (this study). The complex formation and terrane accretion of discrete crustal blocks therefore requires the invocation of a second driving force during the amalgamation of the craton, likely represented by a second subduction zone interacting with the larger, west migrating system. As such, we proposed that the amalgamation of the West African Craton was the product of multiple, diachronous docking or collisional events between discrete juvenile arc terranes. In the case of the Ivory Coast, NW Ghana and western Burkina Faso, the arc terranes may have developed within an oceanic plateau with subsequent collision causing intense structural imbrication of the buoyant oceanic arc lithosphere, analogous to the Belingwe Greenstone Belt and associated terranes in the southern Zimbabwe Craton (Kusky & Kidd, 1992), partially obscuring the early collisional history.

Juvenile crust of the West African Craton represents a unique episode in the geological history of the early Earth, immediately following a hypothesised shut down of magmatism and plate tectonics between 2400 and 2200 Ma (Condie, *et al.*, 2009), and coinciding with the supposed break up of Archaean cratons (Bleeker, 2003). Figure 5.16 shows the spectrum of ϵ Hf_(t) values yielded by magmatic zircons from the study area relative to the global detrital zircon record (Roberts & Spencer, 2015), demonstrating that the juvenile crustal growth in southern Ghana occupies the upper part of the ϵ Hf spectrum for the global record between ca. 2500 and 2000 Ma.

Furthermore, in the global magmatic zircon Hf record, Condie, *et al.* (2011) demonstrated a peak in crustal reworking and a minima in juvenile crust production at 2100 Ma, which significantly contrasts with the crustal record of the study area and the West African Craton. We propose that high juvenile production and preservation recorded in West Africa is the product of the interaction of two northwest-dipping subduction zones. The central/northern subduction zone interacted with an oceanic plateau, as proposed by Boher, *et al.* (1992), associated with the westward migration and cessation of magma attributed to multiple collisional events within the craton and highly varied TTG and magma petrogenetic processes (Goldfarb, *et al.*, 2017; Hirdes & Davis, 2002b; Hirdes, *et al.*, 1996; Parra-Avila, *et al.*, 2017). A smaller subduction zone to the southeast, responsible for the generation of the juvenile crust of southern Ghana (Sefwi, Ashanti and Kibi belts) and the collision along the Sefwi Belt.



Fig. 5.16. Zircon *E*Hf values yielded by this study presented relatively to the global detrital zircon database (Roberts & Spencer, 2015, and references therein), indicating the consistently radiogenic Hf signatures of magmatism in the Sefwi Belt, compared to the broader Hf array of the global record.

Similar crustal evolution and growth models are proposed for the Archaean Belingwe greenstone belt of the Zimbabwe Craton (Kusky & Kidd, 1992) and the Mesoarchaean Rio Maria granite-greenstone terrane of the Amazonian Craton Lateral accretion, age gradients and horizontal shortening are predominantly interpreted as the result of subduction, with several authors proposing such features have characterised accretionary orogens since throughout the majority of Earth's history (Betts, *et al.*, 2011; Card, 1990b; Cawood, *et al.*, 2009; Cawood, *et al.*, 2006; Kusky, 1989; Percival, *et al.*, 1994). The amalgamation of the West African Craton requires multiple subduction zones to elucidate the widespread magmatism, likely interacting with an oceanic plateau representing the recycled proto-crust, as well as the diachronous, distributed collision of multiple terranes. The collisions or intra-terrane sutures do not preserve evidence of modern collisional zones, such as ophiolite sequences or ultra-high pressure metamorphism associated with the subduction of continental crust (Chopin, 2003; Pearce, *et al.*, 1984b), however, evidence of low apparent geothermal gradients, high-pressure—low/mediumtemperature metamorphism and tectonic exhumation of lower orogenic crust has been demonstrated in a number of areas in the craton (this study; Block, *et al.*, 2015; Block, *et al.*, 2016b; Ganne, *et al.*, 2012). We, therefore, propose that the craton represents the ancient expression of a subductioncollision systems, with intermediate characteristics of late Archaean and Phanerozoic accretionary orogenic systems, supporting the concept of the secular evolution and development of modern style plate tectonics during the Palaeoproterozoic (Brown, 2006; Brown, 2007b).

5.7 Conclusions

The Palaeoproterozoic domains of the West African Craton, comprising greenstone belts, large granitegneiss domains and wide meta-sedimentary basins share a number of characteristics with Archaean provinces. The proto-crust was likely extracted from the depleted mantle between ca. 2650 and 2260 Ma, subsequently completely recycled prior to the generation of crustal material preserved in the Sefwi Belt between ca. 2222 Ma and 2080 Ma. Geochemical evidence presented in this study indicates the generation of diverse plutonic suites, including TTG granitoids, which share geochemical affinities with classical volcanic arc rocks whilst displaying striking similarity to Neoarchaean magmatic suites. Major and trace element geochemistry indicates predominantly low-K mafic sources and near ubiquitous positive Pb and Ba anomalies and negative Nb-Ta, P and Ti anomalies, in agreement with models of intra-oceanic subduction-related crustal growth for the craton. Two groups of TTGs as well as dioritic granitoids are emplaced synchronously from contrasting melt sources and melt depths, including partial melting of a young, hot subducting slab, dehydration melting of the metasomatised mantle wedge and shallow partial melting of a mafic source. The subsequent emplacement of quartz monzonites bearing hybrid characteristics and mantle derived magmas and partial melting of preexisting TTGs potentially represents a late extension event, associated with the opening of the Kumasi-Afema Domain between the Sefwi and the Ashanti belts. The emplacement of syn-tectonic two-mica and muscovite granites associated with crustal anatexis is associated with the collision of discrete crustal fragments the cessation of magmatism in southern Ghana. The diversity of magmatic suites documented in the study area resembles that of late Archaean terranes suggesting comparable petrogenetic processes, linked to the global onset of subduction-collision cycles, with Neoarchaean and Palaeoproterozoic terranes indicating the onset occurred locally and diachronously between different cratons (Laurent, et al., 2014).

 $\epsilon H f_{(t)}$ values from analysed samples suggest the Sefwi Greenstone Belt represents an isotopic boundary

within the West African Craton, separating reworked or inherited Archaean crustal material in southeast Ghana from those in the northwest of the Baoulé-Mossi domain. This finding has two implications: (1) that distinct terranes within the craton may potentially be isotopically mapped, allowing us to better understand the crustal architecture; and, (2) that a second Archaean craton lay southeast of the WAC during the Palaeoproterozoic accretion of juvenile crust as indicated by subchondritic ɛHf values in the Kibi Belt, supporting the hypothesis of Petersson, *et al.* (2016). The most significant geodynamic implications of these findings is that we, therefore, have geochemical and isotopic evidence of intraoceanic subduction operating between two Archaean cratons, suggesting that subduction processes were driving tectonic activity in the WAC during the early Palaeoproterozoic. Furthermore, evidence of multiple subduction zones simplifies models that must reconcile rapid generation of multiple volcanic arcs, their diachronous accretion and preservation of such large areas juvenile crust.

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Appendix D – Geochronology of southern Ghana

See pp. 281 - 288 or refer to electronic data files with prefix "Ch5_AppendixD"

Appendix E – Hf standard results

See pp. 289 - 294 or refer to electronic data files with prefix "Ch5_AppendixE"

Appendix F – Hf standards vs accepted values

See pp. 295 - 297 or refer to electronic data files with prefix "Ch5_AppendixF"

Appendix F - Fig. F1: Comparison of measured Hf standards relative to accepted values. Analyses of reference zircons (91500, GJ-1, Mudtank, Plešovice, R33 and MUN 1,3,4) were conducted throughout the session to verify accuracy and ensure reproducibility. These ratios are within the range of TIMS derived values reported for 91500 by Wiedenbeck, et al. (1995) (0.282290 \pm 0.000014), and LA-ICP-MS values for GJ-1 (0.282000 \pm 0.000005) (Morel, et al., 2008), Mudtank (0.282504 \pm 0.000044) (Woodhead & Hergt, 2005), Plešovice (0.282482 \pm 0.000013) (Sláma, et al., 2008), R33 (0.282764 \pm 0.000014) (Fisher, et al., 2014) and doped synthetic zircon standards MUN 1, MUN3 and MUN 4 (0.282140 \pm 0.000006, 0.282135 \pm 0.000005, 0.282141 \pm 0.000006, respectively) (Fisher, et al., 2011).

Chapter VI

Discussion and conclusion



6.1 Discussion

6.1.1 Introduction

This thesis investigated the tectonic, metamorphic and crustal evolution on the volcano-plutonic and volcano-sedimentary rocks exposed in SW Ghana in the Palaeoproterozoic south-eastern portion of the West African Craton. The major aims of this thesis were to:

1) understand the structural and metamorphic evolution of the Sefwi Greenstone Belt;

2) characterise the tectonic setting and driving forces of the Palaeoproterozoic Eburnean Orogeny; and,

3) investigate the geodynamic setting and processes responsible for the formation, amalgamation and preservation of large volumes of juvenile Palaeoproterozoic crust of the West African Craton, and how this reflects global geodynamic changes and the thermal evolution of the Earth.

In order to explore these broad aims, this thesis encompassed a multi-scale and multi-disciplinary approach combining field mapping, structural geology, structural geophysics, metamorphic petrology and *P*-*T* modelling, whole rock major and trace element geochemistry, SHRIMP U-Pb zircon and monazite geochronology, LA-MC-ICP-MS Lu-Hf zircon analysis and LA-ICP-MS U-Pb detrital zircon geochronology. Regional lithological, structural and metamorphic mapping was conducted throughout the 60 000 km² study area, from which we were able to develop a structural and metamorphic framework. This framework, combined with field observations and regional lithological databases, allowed us to interpret regional geophysical datasets to generate new litho-tectonic and metamorphic maps of the Sefwi Greenstone Belt and adjacent volcano-sedimentary domains. In this discussion, we summarise the tectonic and metamorphic history of the study area and potential tectonic models for the Eburnean Orogeny and discuss the insight this provides into Palaeoproterozoic orogenesis. In addition, we discuss the crustal evolution of southern WAC, and how this conceptual model aids our understanding of the evolution of global geodynamics since the Palaeoproterozoic relative to the modern plate tectonic regime. Finally, we highlight the opportunities derived from the thesis for further research in the West African Craton and other Palaeoproterozoic provinces.
6.1.2 Evolution of the West African Craton

6.1.2.1 Tectonic and metamorphic evolution

This thesis provides a new litho-structural and metamorphic map of the Sefwi Greenstone Belt, as well as parts of the Sunyani-Comoé and the Kumasi-Afema domains of southwest Ghana. The work combines a detailed analysis of the metamorphic conditions and thermal regimes recorded by tectono-metamorphic domains within the study area, revealing a far more complex metamorphic history than previously recognised. The thesis presents evidence of an otherwise undocumented migmatitic domain on the north-eastern margin of the Sefwi Belt, tectonically juxtaposed against high-pressure amphibolite facies rocks within the central region of the Sefwi Belt and the low-grade metamorphosed volcano-sedimentary rocks of the Sunyani-Comoé Domain. Documented initial low apparent geothermal gradients (~15°C/km) and clockwise P-T paths are associated with thrust-related crustal thickening, subsequent partial melting in the lower crust and shallow-crustal emplacement of syn-tectonic leucogranites between ca. 2092 and 2081 Ma, parallel to the north-western margin of the Sefwi Belt. Furthermore, high-grade metamorphism and exhumation are attributed to monocyclic orogenesis during Eburnean deformation in the study area, representing the tectonic accretionary phase of the orogen. Combined structural and metamorphic evidence is consistent with collisional orogenesis (England & Thompson, 1986; Thompson & England, 1984).

Late-orogenic exhumation along normal detachments and shear zones during regional transtension resulted in the tectonic juxtaposition of mid-crustal and lower-crustal rocks of the Sefwi Belt with the low-grade, upper-crustal rocks of the Sunyani-Comoé Domain. Constrictional deformation and exhumation of the lower crust are consistent with simple shear-dominated transtension during oblique orogenic extension or collapse (Fossen, *et al.*, 2013), reminiscent of processes recorded in the Early Devonian Scandinavian Caledonides (Dewey, 1988; Krabbendam & Dewey, 1998). Petrological and metamorphic evidence of orogenic collapse includes symplectite textures, supra-solidus isothermal decompression (Duchêne, *et al.*, 1997) and formation of anatectic domes (Rey, *et al.*, 2001; Vanderhaeghe, 2009; Vanderhaeghe & Teyssier, 2001a; Whitney, *et al.*, 2004). This contrasts with the metamorphic record of the study area, which is characterised by concurrent decreases in pressure and temperature indicating cooling during exhumation, juxtaposing both migmatitic and amphibolite facies rocks with greenschist facies supracrustal sequences. Post-collisional extension, exhumation and transcurrent tectonism is favoured over orogenic collapse in

SW Ghana. It is interpreted on the basis of previous studies of elongate granites (Jessell, *et al.*, 2012), coupled with macro-geometries of tectono-metamorphic domains, that D_2 deformation in the Sefwi shear system is characterised by major normal, left-lateral movement. We suggest that extensional strain was localised along the NW margin of the Sefwi Belt due to the presence of a partially molten lower crust along the collisional front, which would have resulted in the mechanical weakening of the lithosphere (Vanderhaeghe, 2012; Vanderhaeghe & Teyssier, 2001a; Vanderhaeghe & Teyssier, 2001b). Our understanding of the driving force of transtension, however, is hindered by the destruction of the Palaeoproterozoic plate margin. Transtension may either be interpreted as the localisation of strain due to a change in boundary conditions, such as rollback of a far-field subduction zone, or, alternatively, it may represent lateral extrusion along a narrow orogenic front (Molnar & Tapponnier, 1975; Sperner, *et al.*, 2002; Thompson, *et al.*, 1997b). In terranes that have undergone extrusion, peak pressures along *P-T-t* paths coincide with peak temperatures (Thompson, *et al.*, 1997a), however, the *P-T-t* paths derived for metamorphic rocks in the study area yield results that are inconclusive in this respect due to poor preservation of prograde assemblages.

As this thesis aimed to explore the evolution and style of tectonic processes during the Palaeoproterozoic, we must also consider the results in light of the alternative tectonic models. In assessing the applicability of a gravitationally-driven tectonic regime, it was established in the introductory chapter of this thesis that sagduction of dense, cold greenstone cover into a mechanically weak felsic lower crust could provide a potential mechanism for the rapid burial of supracrustal rocks (e.g. François, et al., 2014, Ganne, et al. 2014), with corresponding lateral shortening and low apparent geothermal gradients. All of these features are observed in the Sefwi Greenstone Belt, recorded by migmatitic paragneisses of the Chiraa Domain, found in association with voluminous peraluminous granite intrusions, and in garnet amphibolites in the core of Sefwi Belt in the Kukuom-Juaboso Domain. Exhumation of high-grade rocks in the vertical tectonic model requires the diapiric ascent of buoyant granitoids. Whilst this could potentially explain exhumation in the Chiraa Domain, the same can't be said for the garnet amphibolite in the southwest of the Kukuom-Juaboso Domain, which is located within the major Ketesso Shear Zone, distal to any major intrusions. Alternatively, if the entire Sefwi Belt was sagducted, it would require the diapiric emplacement of vast granitoid domains immediately adjacent to the belt. This is not reflected in the crustal architecture of southwest Ghana nor in the isotopic record. Other features observed in the field area that contradict the vertical tectonic model include: dominantly lateral deformation and a near-total absence of steeply downdipping mineral stretching lineations; an asymmetric crustal architecture including structurallycontrolled crust-derived peraluminous granites, orogen-parallel metamorphic gradients and sharp, orogen-oblique lateral metamorphic breaks; and, the presence of vast, low-grade metasedimentary domains either side of the belt. Given the volume and significance of the data that disagrees with the sagduction model, we suggest that it is unlikely to have played a role in the tectonic evolution of the Sefwi Belt.

For the third end-member model, we emphasise that the low apparent geothermal gradients and narrow (<80km) orogenic front documented in this thesis are inconsistent with the hot-orogen model for Precambrian provinces in general (Cagnard, *et al.*, 2011; Cagnard, *et al.*, 2006; Cagnard, *et al.*, 2007; Chardon, *et al.*, 2009; Rey & Houseman, 2006; Rey, *et al.*, 2003), and the West African Craton in particular (Chardon, *et al.*, 2009; Vidal, *et al.*, 2009). The findings of this thesis, therefore, support a plate tectonic regime, within which a nascent form of intra-oceanic subduction generated both arclike magmatism and TTG-like magmatism, as well as sufficient lateral deformation for collisional orogenesis during arc accretion. Furthermore, in the case of the broader geodynamic setting of the formation and amalgamation of the West African Craton, as discussed in Chapter 5, it is possible that there were multiple subduction zones operating simultaneously, likely in association with or adjacent to a Palaeoproterozoic oceanic plateau.

The new geochronological constraints and deformation history presented in this study highlight the diachronous nature of the tectonic evolution of the West African Craton, warning against craton-scale correlation of deformation events, especially on the basis of interpreted orientation of the strain regime. The transition from peak metamorphic temperatures to retrogression and exhumation is constrained by in-situ monazite ages at ca. 2073 Ma, hosted within or aligned with D2 mineral assemblages. We speculate that the preceding terminal collisional event between modern-day SW Ghana and central Ghana/Ivory Coast occurred at ~2100 Ma, which led to the emplacement of syn-tectonic leucogranites along the NW margin of the Sefwi Belt between ca. 2092 and 2081 Ma (Agyei Duodu, *et al.*, 2009; Hirdes, *et al.*, 1992; Hirdes, *et al.*, 2007). This significantly later than the hypothesised collisional event between NW Ghana and SW Burkina Faso at ca. 2130 Ma (Block, *et al.* 2016). The final collision of the Palaeoproterozoic Baoulé-Mossi and Archaean Kénéma-Man domains occurred between ca. 2050 and 2030 Ma (Kouamelan, *et al.*, 1997). These ages indicate lateral variation in the timing of deformation and metamorphism, consistent with progressive accretion and episodic collision of discrete crustal fragments during the Palaeoproterozoic Eburnean Orogeny. This

diachronous nature of deformation and metamorphism raises new questions for future research in the craton. For example, have all episodes of deformation and metamorphism been detected or are there additional internal collisions and what are the implications for the progression of the Eburnean Orogeny?

6.1.2.2 Juvenile crustal growth

The radiogenic Palaeoproterozoic continental crust exposed in Sefwi Belt was derived from a mafic proto-crust extracted from the depleted mantle between 2.26 and 2.65 Ga, indicative of short crustal residence times with minimal reworking of Archaean-aged cratonic material (Ch. 5, this study; Petersson, et al., 2017; Petersson, et al., 2016). The results of coupled geochronological and geochemical studies indicate the coeval production of TTG-like and calc-alkaline magmas, derived from both partial melting of mafic crust at contrasting depths, as well as a LILE-enriched mantle source, with late emplacement of crust-derived felsic magmas. New U-Pb and Lu-Hf analyses indicate the recycling of juvenile mafic crust, with inherited zircon cores dated at ca. 2250 and 2270 Ma. Petrological and geochemical analyses inform us about the petrogenetic processes active during crust formation, including high-pressure partial melting of the subducting slab and low-pressure melting of mafic material generating high SiO2, sodic TTG granitoids with contrasting REE patterns, partial melting of the metasomatised mantle wedge forming dioritic magmas, and late-stage crust-derived granitic plutons. This magmatic evolution is somewhat analogous to Neoarchaean provinces in the Karelian and Kola cratons (Halla, et al., 2009), the Abitibi and Pontiac sub-provinces of the Superior Craton (Feng & Kerrich, 1992), the Rio Maria granite-greenstone terrane of the Carajás Province, Brazil (Almeida, et al., 2011) and the Limpopo Belt and Kaapvaal Craton of southern Africa (Laurent, et al., 2014). New geochronological constraints presented in this study indicate contemporaneous emplacement of magmas from a range of sources within a volcanic arc setting, with late emplacement of magma from shallow crustal sources, interpreted as the product of collision between two crustal blocks. As such, geochemical and isotopic evidences support crustal growth in a subduction setting followed by subsequent arc-arc collision during continental accretion, suturing the arc material of southern Ghana with the arc material of central Ghana and the Ivory Coast, which likely developed within, or in association with, an oceanic plateau. The magmatic evolution of the Sefwi Greenstone Belt falls within the Eoeburnean event of Perrouty, et al. (2012), whilst the proposed collision (this study) corresponds with the Eburnean sensu stricto. I suggest that division of the Eburnean Orogeny into multiple orogenies on the basis of pulses of magmatism must be carefully considered. Whilst

the two peaks of magmatic activity documented by Perrouty, *et al.* (2012) at ca. 2160 and ca. 2090 Ma are likely separated by the deposition of the Kumasi Group, they most clearly demonstrate a change in magma chemistry from calc-alkaline and sodic tonalitic and trondhjemitic magma to more peraluminous, crust-derived magma, thus demonstrating a similar petrogenetic evolution to modern accretionary-collisional orogens.

The proposed collision between southern Ghana and central Ghana/Ivory Coast is contemporaneous with volcanic arc magmatism in the far west of the West African Craton in the Kedougou-Kéniéba Inlier of eastern Senegal (Hirdes & Davis, 2002). As these areas are more than 1000 km apart, we propose that the formation and accretion of juvenile crustal domains in the West African Craton is associated with the interaction between two coeval, west-dipping subduction zones operating simultaneously, and ultimately driving episodic collision between discrete arc terranes. Greenstone belts of southern Ghana are the product of the smaller subduction zone operating independently of a larger, westward migrating subduction zone driving multiple collisional events during interaction with an oceanic plateau in the Ivory Coast, NW Ghana and northern Burkina Faso (Abouchami, et al., 1990; Boher, et al., 1992), and subsequent terminal collision with the Archaean Kénéma-Man Domain at ~2050 to 2030 Ma. This is evident in localised age gradients preserved in both the magmatic and metamorphic record and proposed sutures within the craton (Block, *et al.*, 2016; Eglinger, *et al.*, 2017; Hirdes & Davis, 2002; Kouamelan, et al., 1997; Parra-Avila, et al., 2016; Parra-Avila, et al., 2017; Petersson, et al., 2017; Petersson, et al., 2016; Pitra, et al., 2010). This hypothesis is consistent with the distribution of greenstone belts, the voluminous TTG magma emplacement in the central regions of the southern West African Craton and the diachronous nature of high-grade metamorphism and deformation in the craton. In addition, the rapid formation and accretion of such terranes provides a plausible mechanism for the stabilisation and preservation of such a vast area of juvenile crust.

Finally, the geochemical, isotopic, metamorphic and structural history for SW Ghana presented in this study highlights the striking similarity between the Sefwi Greenstone Belt and the Paleoproterozoic Ile de Cayenne complex of French Guiana. They display coeval juvenile arc-related magmatism between 2174 and 2144 Ma with positive consistently ϵ Nd values, with subsequent crustal thickening and anatexis resulting in the emplacement of small syn-tectonic granite plutons along major shear zones, dated at 2093 ± 8 and 2083 ± 8 Ma (Delor, *et al.*, 2003; Vanderhaeghe, *et al.*, 1998). The final correlation between the two terranes comes from the timing of crustal extension and formation of granulite facies metamorphism is in the Guiana Shield at 2070 to 2060 Ma (Delor, *et al.*, 2001),

which we interpret to be a lateral continuation of the D2 deformation and metamorphism recorded on the northwest margin of the Sefwi Belt. This suggests that the Sefwi Belt and the Ile de Cayenne were contiguous during the Palaeoproterozoic Eburnean-Transamazonian accretion/collision orogeny. This hypothesis is supported also by the formation of granulitic domains during late-orogenic crustal extension in the Ile de Cayenne (Delor, *et al.*, 2001), potentially representing the spatial and temporal continuation of the constrictional deformation event documented in the Sefwi Belt in this study.

6.1.3 Implications for Palaeoproterozoic geodynamics and orogenesis

Controversy surrounding the geodynamic setting of the formation of juvenile crust of the West African Craton is borne of the relatively limited geochronological constraints on the timing of deformation and metamorphism, as well as ambiguous or often obscured, field relationships between greenstone belts, granite-gneiss domains, and sedimentary basins, a problem also frequently encountered in Archaean terranes (Condie, 1981; de Wit, 1998). We conclude that the magmatic record of the West African Craton represents an ancient example of the subduction-collision tectonics, with a crustal evolution reminiscent of a number of Neoarchaean provinces formed at the advent of "modern-subduction" processes (Laurent, *et al.*, 2014; Martin & Moyen, 2002). It represents the amalgamation of arc terranes forming in an intra-oceanic setting. We emphasise that the original distribution of the terranes may have been thousands of kilometres apart, with their amalgamation requiring the closure of multiple oceans over the ~ 200 Myr evolution of the Eburnean-Transamazonian Orogeny, potentially analogous to the Miocene geodynamics of the Maramuni arc of Papua New Guinea, and the South Pacific (Holm, *et al.*, 2016; Holm, *et al.*, 2015; Schellart, *et al.*, 2006).

The Neoarchaean to Palaeoproterozoic period coincides with a number of significant changes in the Earth, including changes in the tectonic style, secular evolution recorded in the metamorphic record and evolving chemistry of magmatic rocks (e.g. Brown, 2006; Brown, 2007; Dhuime, *et al.*, 2012; Keller & Schoene, 2012; Laurent, *et al.*, 2014). This is a unique period prior to the Neoproterozoic appearance of blueschist facies or ultra-high-pressure (UHP) metamorphism and tectonic indicators, such as ophiolite sequences, attributed to the modern plate tectonic regime (Stern, 2005; Stern, 2007). It is widely accepted that the ambient upper-mantle temperatures were hotter in the past (e.g. Herzberg, *et al.*, 2010; Labrosse & Jaupart, 2007), which likely affected the tectonic style and the rheology of the lithosphere (Burov & Yamato, 2008; Burov, 2011; Sizova, *et al.*, 2010; van Hunen & van den Berg, 2008). Rey and Coltice (2008) propose that the secular cooling of the Earth over the

Archaean-Proterozoic transition is coupled with a cooling and strengthening of continental crust, after which, the continental crust should theoretically support thrust-related crustal thickening and elevated topography. This study supports a growing body of research in the West African Craton that demonstrates thrust-related crustal thickening and structurally-controlled exhumation in a relatively strong lithosphere (Block, et al., 2016; Feybesse, et al., 2006; Hirdes, et al., 2007). The tectonic style and orogenic model proposed for SW Ghana bears a number of similarities to collisional orogenesis documented in the Archaean Barberton terrain (Diener, et al., 2013; Moyen, et al., 2006) and the Archaean Abitibi Sub-province (Daigneault, et al., 2002; Mueller, et al., 1996) and the Early Devonian Western Gneiss Region of the Scandinavian Caledonides (Dewey, 1988; Krabbendam & Dewey, 1998). In the Palaeoproterozoic rocks of southern Ghana, the absence of UHP rocks, akin to those found that characterise the Scandinavian Caledonides, highlights a conundrum in Neoarchaean to Mesoproterozoic terranes eloquently elucidated by Brown (2007). Brown poses the question as to whether the absence of such rocks reflects: a) the inactivity of deep subduction and generation of UHP rocks at this time in the Earth's history, or, b) the inability of the early Earth lithosphere to exhume rocks from such depths, on the basis of elevated geotherms resulting in a rheologically weaker lithospheric plate. We suggest that the crustal record of the West African Craton represents nascent subductionrelated plate tectonics, tempered by elevated lithospheric temperatures during the Palaeoproterozoic relative to today. This is supported by coeval generation of slab-derived high-pressure TTGs, which requires elevated geotherms in the downgoing slab, and calc-alkaline dioritic magmas characteristic of a modern arc terrane. Furthermore, the metamorphic record reveals cold apparent geothermal gradients and HP-LT and HP-MT metamorphic assemblages similar to modern collisional orogens (this study; Block, et al., 2015; Ganne, et al., 2012). The absence of UHP assemblages likely reflects the inability of the slightly warmer and weaker Palaeoproterozoic crust to generate or exhume such assemblages, supporting the hypothesis of Sizova, et al. (2014), who propose a secular evolution of collisional orogenesis during the Proterozoic. The hybrid Archaean-Phanerozoic characteristics of subduction-related magmatism and subduction-collision-related metamorphism suggests that the evolution of geodynamic processes may have been a smooth progression, with gradual adoption of modern characteristics and the appearance of classic tectonic indicators occurring in the Palaeoand Mesoproterozoic. If this is the case, Palaeoproterozoic terranes around the world may provide the missing link between archaic and modern tectonic regimes, thus warranting further geological research.

6.1.4 Recommendations for future research

Based on the findings of this thesis, more thorough, suite-specific geochemical and isotopic analysis would further improve our understanding of the temporal evolution of magmatism and petrogenetic processes within the Sefwi Greenstone Belt. The expansion of the TerraneChron[®] program (Parra-Avila, 2015, and references therein) may facilitate wider scale isotopic mapping of the craton and to examine the possibility of variance in the age of the 2300– 2200 Ma proto-crust being recycled in the WAC, as evidenced by the limited inherited zircon data. Alternatively, continued systematic, regional Lu-Hf, Sm-Nd and O isotope analysis may provide greater insight into the timing of mantle contribution and reworking of existing crust (e.g. Hawkesworth & Kemp, 2006). In addition, this may help better resolve questions regarding the mono- or polycyclic history of the Eburnean Orogeny by informing us about the true basement of the West African Craton.

Continued high-resolution mapping within the belt at a domain-scale may provide additional insight into deformation within each domain as well as information regarding the tectonic nature of domain boundaries. Such studies could be complemented by structural mapping and geochronological studies conducted across the width of the Sunyani-Comoé and Kumasi-Afema domains to better constrain the relative timing of volcano-sedimentary deposition and deformation of each domain, as well as potentially elucidating the stratigraphic and tectonic relationships with the Bui and Ashanti greenstone belts. Lu-Hf dating of garnet growth from throughout the Sefwi Belt would elucidate the timing of HP metamorphism and D1 deformation for better correlation between domains within the belt. Additional geochronological constraints on the timing of high-grade metamorphism and mineralisation in both the Sefwi Belt and Ashanti Belt would greatly improve our understanding of the relative timing of deformation between the two greenstone belts, and their spatial association during the Eburnean Orogeny.

It is recommended that future research focuses on continued high-resolution mapping of high-grade terranes, particularly along shear zones within central and south-western Ivory Coast, in order to investigate other potential sutures within the craton. This may be supplemented with isotopic studies to investigate other potential isotopic boundaries in order to better understand the crustal architecture of the West African Craton. Additional analysis of O-isotope systematics in zircons from the WAC would greatly complement Hf data by providing insight into the geodynamic setting of crust formation as well as information on the interaction of mantle-derived melts and supracrustal material exposed to near-surface processes, such as hydrothermal alteration (Valley, 2003). Structural studies should be complemented by quantitative calculations of metamorphic P-T paths and direct, high-precision dating of deformation and metamorphism. There should be a degree of caution when considering correlating craton-wide magmatic and tectonic events, as this leads to an over-simplification of stratigraphic and tectonic relationships and a loss of detail, which is detrimental to the understanding of the growth and amalgamation of the craton. Furthermore, we suggest that the Eoeburnean vs Eburnean relationship be reassessed in a more detailed, spatial context in order to better understand lateral age gradients and tectono-thermal relationships between adjacent terranes.

6.2 Conclusion

The Palaeoproterozoic crust of the West African Craton provides a natural laboratory to explore the geodynamic and orogenic links between the Archaean and the Proterozoic, preserving a unique period of juvenile crustal growth in the global rock record. The magmatic, tectonic and metamorphic evolution of the Palaeoproterozoic West African Craton reflects a nascent horizontal plate tectonic regime, highlighting the progressive changes in the geochemical and metamorphic record.

Within this thesis, I established a new magmatic framework for the Palaeoproterozoic Sefwi Greenstone Belt, which indicated juvenile crust production and collision of subduction-related intraoceanic arcs. Striking similarities exist between the petrogenetic processes and magmatic evolution of SW Ghana and Neoarchaean terranes as well as modern arc systems. The geochemical and isotopic record of the region reflects the gradual adoption of modern petrogenetic and crustal processes. This is supported by the tectonic and metamorphic history of the region, characterised by initial thrust-related crustal thickening and generation of HP-MT metamorphic conditions which is interpreted to represent the collision of arc terranes. The subsequent tectonic exhumation of the lower crust is evidence of strengthening of the lithosphere, inconsistent with the proposed *hot-orogen* model for the early Proterozoic. The implication of this proposed geodynamic and tectonic evolution for the West African Craton informs our understanding of the global tectonic regime and secular cooling of the early Earth.

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Appendices

Appendix A - Supplementary data I - IV

The following information was used to constrain the new litho-tectonic maps for southwest Ghana.



Fig. A1SI. Map of displaying survey coverage of airborne magnetic surveys. The residual magnetic intensity (*RMI*) was reduced to the pole (*RTP*) to remove any asymmetry of magnetic anomalies caused by the inclination of the Earth's magnetic field. The *RTP* images highlight the *N*-S artefacts commonly produced by the *RTP* transformation when performed in areas of low latitude (Li, 2008; MacLeod, et al., 1993).

Data	Magnetic	Radiometric	Magnetic	Radiometric	Magnetic	Radiometric	Magnetic
Country	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	lvory Coast
Survey	A1	A1	B,D,E	B,D,E	С	С	F
Year	1997	1997	1999	1999	1984	1984	1976
Flight height (m)	70	70	80	80	80	80	152
Line direction (°)	135	135	B, E: 90 D: 000	B, E: 90 D: 000	90	90	000
Line spacing (m)	400	400	400	400	200	200	500
Image pixel size (m)	100 x 100	100 x 100	100 x 100	100 x 100	100 x 100	100 x 100	125 x 125

Table A1SI. Individual airborne magnetic survey details.



Bouguer gravity anomaly grid (mGal)

Fig. A2SI. Bouguer gravity anomaly grid highlighting the dominant northeast trend in the regional architectures, with gravity highs associated with the central Sefwi Greenstone Belt and the Ashanti Greenstone Belt in the southeast. The Kumasi-Afema Domain between the two belts is associated with a 40km-wde intermediate gravity anomaly, and intermediate to low-gravity anomalies associated with the Sunyani-Comoé Domain.



Negative reduced to the equator (RTE) of residual magnetic intensity (RMI) grid (nT) with greyscale first vertical derivative (1VD) overlay

Fig. A3SI. Negative reduced to the equator (RTE) of residual magnetic intensity (RMI) grid in colour using to interpret litho-tectonic domains and some lithology distribution with greyscale first vertical derivative (1VD) overlay to highlight magnetic fabrics and major lineaments.



Fig. A4SI. Composite image displaying a greyscale scale digital elevation model with 90m resolution (Shuttle Radar Topographic Mission, SRTM, 2000) with an overlay of a ternary image of gamma-ray signal from three channels, K, eTh, and eU, highlighting the partial coverage of radiometric data of the study area.



Appendix B – Magnetic susceptibility histograms

Fig. B1. Compilation of measured magnetic susceptibility values for lithologies in the study area. Measurements were made using an RT-1 Magnetic Susceptibility Meter.

Appendix C – Pseudosection data files

Perplex data files for pseudosection calculation.

Refer to electronic data files with prefix "Ch4_AppendixC_Perplex"

Appendix D – Geochronology of southern Ghana

outhern Ghana					Sample							
Age (Ma)	10	Lat	Lon	Country	name	Litho	Domain	Technique	Mineral	Locality	Ref McFodooo of of	Interp
2270				Ghana	SB023		Sw	U-Pb SHRIMP	Zr		unpublished	_
2250				Ghana	SB023		Sw	U-Pb SHRIMP	Zr		McFarlane et al. unpublished	-
2197	10	6.477	-1.133	Ghana	GH390		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1996	-
2156	80	6.477	-1.133	Ghana	GH390		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1998	-
2128	4	6.477	-1.133	Ghana	GH390		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1998	-
2156	e	6.263	-1.6069	Ghana	GH388		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1996	-
2123	7	6.263	-1.6069	Ghana	GH388		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1998	
2144	6 2	6.263	-1.6069	Ghana	GH1014		As	Pb-Pb (dissolution)	Zr		Oberthür et al. 1998	-
2159	12	6.263	-1.6069	Ghana	GH789		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1998	-
2187	t	6.263	-1.6069	Ghana	GH1014		As	Pb-Pb (dissolution)	Zr		Oberthür et al., 1996	-
usive												
2229	4	5.622	-0.33005	Ghana	PK 105	Biotite-hornblende granite	K-S	U-Pb SIMS	Zr	West Accra	Petersson et al. 2016	υ
2219	9	6.023967	-0.475367	Ghana	ASGH030A	Biotite-homblende granite	K-S	U-Pb SIMS	Zr	Badu Kibi belt	Petersson et al. 2017	O
2200	4	5.98	-0.5233	Ghana	G113B	Granodiorite	K-S	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	υ
2189	9	6.124817	-0.2325	Ghana	ASGH028A	Biotite-hornblende granite	K-S	U-Pb SIMS	Zr	Nkurakan Kibi belt	Petersson et al. 2017	ပ
2187	1	5.11315	-1.51348	Ghana		Gneiss	S	dq-dq	Zr		Loh et al., 1999	O
2182	2	5.49974	-2.98864	Cote d'Ivoire			Su	U-Pb TIMS	Zr		Agyei Duodu et al., 2009	o
2179	2	6.130124	-2.551952	Ghana	6668	Granitoid	Sw	Pb-Pb (dissolution)	Zr		Hirdes et al., 1992	C
2174	3	4.9555	-1.6629	Ghana	GH852	Granitoid	Sc	Pb-Pb (dissolution)	Zr		Oberthür et al., 1998	O
2174	9	5.811	-0.34975	Ghana	PK102	Biotite-hornblende granite	K-S	U-Pb SIMS	Zr	Nsawam	Petersson et al. 2016	υ
2173	2	4.793483	-1.94555	Ghana	ASGH007A	Homblende tonalite	۲	U-Pb SIMS	Zr	Dixcove	Petersson et al. 2016	υ
2172	2	4.79161	-1.94793	Ghana	Dixcove	Gneiss	ő	U-Pb	Zr		Opare Addo, 1992	U
2172	2	4.834242	-1.908936	Ghana	6427	Granitoid	ő	Pb-Pb (dissolution)	Zr		Hirdes et al., 1992	O
2167	4	6.186317	-2.426683	Ghana	ASGH019A	Biotite-homblende granite	Sw	U-Pb SIMS	Zr	Wiawso	Petersson et al. 2017	O
2165	6	5.39542	-0.58077	Ghana		Gneiss	K-S	U-Pb	Zr		Opare Addo, 1992	υ
2164	8	5.475	-1.865	Ghana	AWADI	Diorite	As	U-Pb SIMS	Zr		White et al. 2014	O
2162	٠	5.840333	-3.202833	Cote d'Ivoire	GB2293	Oz diorite	Sw	Pb-Pb (dissolution)	Z		Hirdes et al., 2007	U
2161	7	5.403	-3.128	Ghana	Sefwi 227	Monzonite	Sw	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	υ
2161	4	5.583	-3.426	Ghana	Setwi 238	Monzonite	Sw	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	ပ
2161	7			Ghana	Setwi 225	Diorite	Sw	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	U
2159	4	4.83	-2.16	Ghana	BPD61	Granodiorite	As	U-Pb TIMS	Zr		Attoh et al., 2006	O
2159	4	7.093	-2.4098	Ghana	Sefwi D14	Granitoid	Sw	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	O
2159	4	5.43224	-2.7751	Cote d'Ivoire			Ku	U-Pb TIMS	Zr		Agyei Duodu et al., 2009	υ
2158	00	5.166667	-1.9	Ghana			As	U-Pb SHRIMP	Zr		Parra-etal_unpublished_W	υ
2158	ŝ	5.517	-0.445	Ghana	G1d	Monzonite arthogneiss	K-S	Pb-Pb (evaporation)	Zr		Feybesse et al., 2006	υ
2154	5	6.268078	-2.481873	Ghana	SB248	Trondjhemite	Sw	U-Pb SHRIM(Zr		McFarlane et al unpublishe	υ
2148	ŝ	5.7655	-3.184833	Cote d'Ivoire	GB2165	Granite	Sw	Pb-Pb (dissolution)	Zr		Hirdes et al., 2007	U
2143	10	6.97169	-2.37792	Ghana	SB092	Tonalite	Sw	U-Pb SHRIMP	Zr		McFarlane et al unpublishe	O
2139	9	5.34355	-0.616533	Ghana	ASGH001A	Granite	K-S	U-Pb SIMS	Zr	Kibi-Winneba	Petersson et al. 2017	U
2139	5	5.436417	-0.472867	Ghana	PK103	Biotite-homblende granite	K-S	U-Pb SIMS	Zr	Gomoa Fetteh	Petersson et al. 2016	o

0	0	O	U	U	U	O	U	U	0	O	υ	U	U	O	υ	U	O	0	U	0	0	0	U	o	U	0	0	0	0	U		U	0	0	υ	U	U	0	O	O			Z	N
Petersson et al. 2017	Petersson et al. 2017	Hirdes et al., 2007	Adadey et al., 2009	Amponsah, P. MSc 2011	McFarlane et al unpublishe	Amponsah, P. MSc 2011	Agyei Duodu et al., 2009	Petersson et al. 2017	Petersson et al. 2016	Petersson et al. 2016	Petersson et al. 2017	Hirdes et al., 1992	Agyei Duodu et al., 2009	Adadey et al., 2009	Adadey et al., 2009	Oberthür et al., 1998	Agyei Duodu et al., 2009	Oberthür et al., 1998	Oberthür et al. 1998	Adadey et al., 2009	Agyel Duodu et al., 2009	Oberthür et al. 1998	Petersson et al. 2016	Adadey et al., 2009	Davis et al., 1994	Agyei Duodu et al., 2009	Hirdes et al., 1992	Hirdes et al., 2007	Agyei Duodu et al., 2009	Agyei Duodu et al., 2009		Loh et al., 1999	Parra-etal_unpublished_W	Hirdes et al., 1998	Hirdes et al., 2007	White et al. 2014	Hirdes et al., 2007	Delor et al., 1992.	Loh et al., 1999	McFarlane et al unpublishe	Adadey et al., 2009		Hirdes et al., 2007	Oberthür et al., 1998
Sasi Kibi belt	Abcabo Kibi belt							Bole	Amasaman	Cape Coast	Maluwe												Sunyani																					
Zr	Zr	Zr	Zr	Zr	Zr	Z	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Mnz	Zr	Mnz	Mnz/Ttn	Zr	Zr	Ttn	Zr	Zr	Zr	Zr	Zr	Zr/Mnz	۶ł	Zr		Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr		đ	R
U-Pb SIMS	U-Pb SIMS	Pb-Pb (dissolution)	U-Pb SHRIMP	0-Pb	U-Pb SHRIMP	u-Pb	U-Pb	U-Pb SIMS	U-Pb SIMS	U-Pb SIMS	U-Pb SIMS	Pb-Pb (dissolution)	U-Pb	U-Pb SHRIMP	U-Pb SHRIMP	Pb-Pb (dissolution)	U-Pb	Pb-Pb (dissolution)	Pb-Pb (dissolution)	U-Pb SHRIMP	U-Pb	Pb-Pb (dissolution)	U-Pb SIMS	U-Pb SHRIMP	Pb-Pb (evaporation)	U-Pb	Pb-Pb (dissolution)	Pb-Pb (dissolution)	U-Pb	u-Pb		Pb-Pb	U-Pb SHRIMP	Pb-Pb (dissolution)	Pb-Pb (dissolution)	SMIS 44-N	Pb-Pb (evaporation)	U-Pb monazircon	Pb-Pb	U-Pb SHRIMP	U-Pb SHRIMP		Pb-Pb (dissolution)	Pb-Pb (dissolution)
K-S	K-S	Sw	Ku	Sw	Su	Sw	K-S	BBD	K-S	0-0	N	Ku	K-S	Ku	Ки	As	K-S	Ки	As	As	Sc	As	Su		As	Su	Su	Su	Cc	Cc		As	As	Sw	Sw	As	Sw	Sw	As	Sw	Ku		Sw	As
Biotite-homblende granite	Biotite-homblende granite	Qz monzonite	Granite	Granitoid	Granite	Granitoid	Gneiss	Biotite granite	Biotite-homblende granite	Two mica granodiorite	Two mica granodiorite	Biotite tonalite	Granite	Granite	Granite	Granite	Granodiorite	Granite	Granitoid	Diorite	Granite	Granodiorite	Muscovite biotite granite	Granite	Monzonite	Granite	Ms granitoid	Granite 2 micas	Pegmatite	Pegmatite		Early Birimian volcanics		Rhyolite	Intermediate volcanoclastite	Volcanoclastite	Felsic volcanic	Rhyolite	Volcanic rock	Dacite	Andesite		Qz diorite	Granite
ASGH032A	ASGH034A	GB2145	AK2091	ASU002	SB023	NF002		ASGH048A	PK101	ASGH003A	ASGH046A	6899A		AK5216	AK4018	3H789, GH101		GH794	3H1017, GH38	AK1334		GH390	ASGH022A/C	AK1249	HCP1-3		8909A	CB2324						H285		AWABV				SB173B	AK4124		GB2293	GH789
Ghana	Ghana	Cote d'Ivoire	Ghana	Ghana	2 Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Chana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Ghana	Cote d'Ivoire	Ghana	Ghana		Ghana	Ghana	Ghana	Cote d'Ivoire	Ghana	Cote d'Ivoire	Cote d'Ivoire	Ghana	Ghana	Ghana		Cote d'Ivoire	Ghana
-0.60235	-0.553917	-3,157833	-2.444778	-2.7672	-2.68039672	-2.7688	-0.62127	-2.441933	-0.271167	-0.6138	-2 2237	-2.216191	-0.623724	-2.17879	-2.50108	-1,6069	-0.424909	-1.902	-1.6069	-1.99924	-1.3073	-1.133	-2.1836	-2.22026	-1.34664	-2.65483	-2.536375	-3.181167	-0.96981	-1.16126		607700	-1.7435	-2.0306	-3.032333	-1.865	-3 2005	-3.115236	607700	-2.290199	-2.44706			-1 6069
5.96705	5.944083	5.7695	5.980056	6.7055	6.71046969	6.6847	5.63842	9.439183	5.712167	5.343983	8.58225	6.487087	5.33529	5,9815	5.65248	6.263	5.57583	5.916	6.263	5.52658	5.75501	6.477	7.4807	5.67238	5.145756	7.06817	7.095005	5,9655	5,48501	5.15365		526300	5.4606	6.9075	5.672167	5.475	5.621833	5.580471	526300	7.38019	5.58705			6.263
7	Q	2	თ	1	7	ŋ	4	ß	12	18	9	ы	-	19	18	ო	٣	ი	2	13	٣	2	2	44	÷	-	+	-	ę	٣		2	Q	٣	2	Ø	N	1	9	9	24		ო	7
2138	2137	2137	2136	2136	2136	2135	2132	2130	2126	2125	2120	2116	2113	2112	2108	2106	2106	2105	2104	2102	2102	2097	2093	2090	2090	2068	2088	2081	2080	2072	Extrusive	2266	2193	2189	2180	2178	2166	2164	2162	2158	2142	Metamorphic	2144	2098

2092	ო	4.9555	-1.6629	Ghana	GH852	Granitoid	ö	Pb-Pb (dissolution)	Ē	Oberthür et al., 1998	⋝
2073				Ghana	SB398	Paragneiss	Sw	U-Pb SHRIMP	Mnz	McFarlane et al. unpublish	⋝
				Ghana	Sb389	Paragneiss	Sw	U-Pb SHRIMP	Mnz	McFarlane et al. unpublish	⋝
Detrital grain											
2245	4.0	5.297506	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2194	3.0	5.358142	-1.94285	Ghana	2069		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2189	2.0			Cote d'Ivoire	GB0027		Sw	Pb-Pb (dissolution)	Zr	Hirdes et al., 2007	0
2185	2.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	G
2184	3.0	5.81708	-2.016435	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2184	3.0	5.830652	-2.00652	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	0
2183	3.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2180	4.0	5.297508	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2179	3.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2174	5.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2173	4.0	5.358142	-1.94285	Ghana	2069		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	\$
2172	4.0	5.297506	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2172	5.0	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2170	15.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	\$
2169	11.0	5.297506	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	\$
2168	4.0	5.297506	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2168	3.0	5.830652	-2.00652	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	0
2167	3.0	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2167	3.0	5.830652	-2.00652	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2164	7.0	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	0
2163	31.0	5.358142	-1.94285	Ghana	2069		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2161	3.0	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2161	2.0			Cote d'Ivoire	GB0024			Pb-Pb (dissolution)	Zr	Hirdes et al., 2007	00
2159	4.0	5.297506	-1.991515	Ghana	TK013		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	0
2158	3.0	5.342788	-1.987351	Ghana	TK001		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2155	5.0	5.358142	-1.94285	Ghana	2069		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	\$
2155	5.0	5.830652	-2.00652	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2152.5	19.5	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	ŝ
2135	5.0	5.81708	-2.016435	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	0
2135	5.0	5.81708	-2.016435	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2135	5.0	5.830652	-2.00652	Ghana	HUN1-3		Ku	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2133	3.0	5.297636	-1.990735	Ghana	Wta34		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2132	3.0	5.358142	-1.94285	Ghana	2069		As	Pb-Pb (dissolution)	Zr	Davis et al., 1994	S
2129	2.0			Cote d'Ivoire	GB0024			Pb-Pb (dissolution)	Zr	Hirdes et al., 2007	ŝ
2129	2.0			Cote d'Ivoire	GB0027			Pb-Pb (dissolution)	Zr	Hirdes et al., 2007	S
2132	2.8	5.337	-1.998	Ghana			As	U-Pb	Zr	Hirdes and Nunco., 1994	S
2132.6	3.4	5.337	-1.998	Ghana			As	U-Pb	Zr	Hirdes and Nunco., 1994	ŝ
2155.0	4,0	5.337	-1.998	Ghana			As	u-Pb	Zr	Hirdes and Nunso., 1994	\$
2158.0	4.2	5.337	-1.998	Ghana			As	U-Pb	Zr	Hirdes and Nunco., 1994	S
2158.2	3.1	5.337	-1.998	Ghana			As	0-Pb	Zr	Hirdes and Nunoo., 1994	0

S	S	S	ŝ	S	S	S	S	S	S	S	S	ŝ	ŝ	S	S	0	S	s	S	ŝ	S	S	S	ŝ	S	S	တ	ŝ	ŝ	S	S	S	S	S	ŝ	S	S	S	ŝ	S	S	0	ŝ	S
Hirdes and Nunoo., 1994	Hirdes and Nunco., 1994	Hirdes and Nunco., 1994	Hirdes and Nunoo., 1994	Hirdes and Nunco., 1994	Hirdes and Nunco., 1994	Hirdes and Nunco., 1994	Hirdes and Nunoo., 1994	Hirdes and Nunoo., 1994	Hirdes and Nunco., 1994	Hirdes and Nunco., 1994	Hirdes and Nunoo., 1994	Loh et al., 1999	Pigois et al., 2003																															
Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr	Zr														
U-Pb	U-Pb	U-Pb	U-Pb	d-Pb	0-Pb	0-Pb	d-Pb	dq-U	0-Pb	U-Pb	d-Pb	94-N	U-Pb	U-Pb	Pb-Pb	U-Pb SHRIMP	U-Pb SHRIMP	U-Pb SHRIMP	U-Pb SHRIMP	U-P5 SHRIMP	U-Pb SHRIMP																							
As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As	As														
														22																											-			
Ghana	Ghana	Ghana	Ghane	Ghane	Ghana	Ghana	Chane	Ghana	Ghana	Ghane	Chane	Ghane	Ghana	Ghane	Ghane	Ghane	Ghana	Ghane	Ghana	Ghana	Ghana	Ghane	Ghana	Ghana	Ghane	Ghane	Ghana	Ghana	Ghane	Ghane	Ghana	Ghana	Ghana	Ghane	Ghana	Ghana	Ghana	Ghana	Ghana	Ghane	Ghane	Chane	Ghana	Ghana
-1.998	-1 998	-1.998	-1,998	-1.998	-1.998	-1,998	-1,998	-1.998	-1.998	-1.998	-1.998	-1.998	-1.998	-2.030	-2.030	-2.030	-2.030	-2.030	-2 030	-2.030	-2.030	-2.030	-2.030	-2.030	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1,862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1,862	-1.862	-1.862
5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	5.337	4.761	4.761	4.761	4.761	4.761	4.761	4.761	4.761	4.761	4.761	4.761	5.517	5.517	5.617	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517
4.0	3.4	7.1	2.8	3.7	5.1	3.5	3.8	5.8	2.8	42	3.2	2.4	2.9	3.5	4.8	3.1	2.2	2.0	3.1	1.5	1.4	3.8	7.3	2	17	17	26	19	16	61	76	10	19	18	30	22	27	36	15	29	22	16	14	F1
2158.5	2160.5	2164.0	2167.1	2168.1	2171.5	2172.0	2173.5	2174.1	2179.0	2180.0	2184.4	2184.5	2194.2	2244.6	2152.6	2161.9	2166.6	2166.7	2168.2	2182.3	2256.7	2262.1	2264.2	2265.5	2090	2095	2100	2103	2104	2104	2105	2109	2109	2109	2111	2111	2113	2113	2114	2114	2115	2116	2116	2118

60	ω	ŝ	90	S	S	ŝ	ŝ	ŝ	S	Ŋ	S	\$	ŝ	ŋ	ŝ	ŝ	ŝ	()	ŝ	Q	ŝ	ŝ	S	ŝ	00	ŋ	S	ŝ	Ø	S	(n)	Q	S	\$	S	S	ŝ	ŝ	ŝ	ŝ	ŝ	ω	Ø	60
Pigois et al., 2003	Pigois et al , 2003	Pigois et al., 2003																																										
Zr	C	Zr	Z	Zr	Z	Zr	Z	Zr	Zr	Zr	Zr	Zr	Zr	Z	Zr	Z																												
U-Pb SHRIMP																																												
As																																												
Ghana																																												
-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1 862	-1.862	-1.862	-1.862	-1.862	-1.862	-1 862	-1 862	-1.862	-1.862	-1.862	-1.862	-1.862	-1 862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1,862	-1,862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862	-1.862
5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5,517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5,517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517
27	12	24	16	18	თ	15	18	17	16	16	14	£.	23	12	17	41	18	18	11	15	თ	12	32	6	30	13	16	13	19	16	16	23	13	21	ø	17	52	23	19	23	16	11	44	19
2119	2119	2119	2119	2122	2122	2122	2123	2123	2123	2124	2124	2124	2125	2126	2128	2131	2131	2131	2133	2134	2134	2135	2135	2136	2140	2142	2142	2142	2142	2143	2143	2143	2144	2144	2145	2145	2146	2146	2146	2147	2148	2149	2151	2153

s	Ø	ŝ	s	s	S	s	ŝ	Ø	IJ	ŝ	ø	ŝ	Ø	Ø	ŝ	Ś	s	S	s	s	ø	S	S	ø	S	s	S	s	S	s	ø	ŝ	s	s	S	Ø	S	so	s	S	S	ŝ	Ø	S
Pigois et al., 2003																																												
Zr																																												
U-Pb SHRIMP																																												
As																																												
ារាង	na	ina	រោង	Ina	ua.	Lia	ពេង	III	ua.	ua.	ina.	Ina	Lia.	10.8	Ua.	lina.	កេត	Ina	Lia -	ហេង	ina.	ina.	Ina	na	INB	ILIA	113	ua.	Ina	11.3	Ina	na	ma	Ina	INA	na	ina.	IL B	ILB	IUB	li a	ina.	ពេង	na
2 Ghe	2 Ghs	2 Gha	2 Ghe	2 Ghe	2 Gha	2 Gha	2 Gha	2 Gha	2 Ghe	2 Gha	2 Gha	2 Ghe	2 Ghe	2 Ghe	2 Ghá	2 Gha	2 Ghe	2 Ghs	2 Gha	2 Gha	2 Che	2 Ghe	2 Gha	2 Gha	2 Ghe	2 Gha	2 Ghe	2 Gha	2 Gha	2 Gha	2 Ghs	2 Gha	2 Gha	2 Ghe	2 Ghe	2 Ghs	2 Gha	2 Gha	2 Gha	2 Ghe	2 Gha	2 Gha	2 Ghe	2 Gh
-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1,86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1,86	-1.86	-1.86	-1,86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1.86	-1,86	-1.86	-1.86	-1.86	-1.86	-1.86
5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517	5.517
10	22	24	21	17	g	31	28	22	25	23	19	21	24	18	44	12	20	19	4	23	19	20	31	21	44	20	22	17	4	12	17	12	10	23	Ę	22	23	19	24	14	25	15	19	29
2155	2155	2155	2156	2156	2158	2159	2159	2159	2162	2163	2163	2164	2164	2165	2165	2168	2168	2169	2173	2174	2174	2176	2176	2179	2180	2180	2180	2182	2182	2184	2186	2187	2188	2188	2191	2198	2198	2200	2205	2211	2222	2223	2236	2242

S		
Pigois et al., 2003		
Zr		
U-Pb SHRIMP		
As		
Ghana		
-1.862		
5.517		
21		
2257		

Appendix E – Hf standard results

	If180Hf1	7m	1.94347	1.94069	1.94098	1.9397	1.93886	1.93822	1.93912	1.93728	1.93705	1.9366	1.93677	1.93676	1.93662	1.936	1.93633	1.93652	1.93649	1.93557	1.93508	1.94442	1.94052	1.94074	1.93939	1.93961	1.9373	1.93757	1.93681	1.93696	1.93664	1.93632	1.93593	1.93616	1.93605	1.93641	1.93573	1.93601	1.93639	1.93603	1.93558
If179Hf1	7m_Int2 H		160000.0	0.000089	0.00011	0.000071	0.00017	0.000063	0.00011	0.000079	0.00011	0.000092	0.00011	0.000093	0.000074	0.000079	0.000097	0.000087	0.000071	0.0001	0.000077	0.000085	0.00013	0.00011	0.0001	0.00014	0.000079	0.000083	0.000074	0.000076	0.000089	0.000063	0.000079	0.000083	0.000094	0.000072	0.00001	0.000096	0.00016	0.000088	0.000066
-	Hf179Hf1 7	7m 5	0.747103	0.746392	0.74646	0.74622	0.74591	0.745776	0.74596	0.74543	0.74551	0.745372	0.7454	0.745391	0.745377	0.745193	0.745282	0.745352	0.745343	0.74512	0.745025	0.74738	0.74633	0.7464	0.7461	0.7461	0.745538	0.745544	0.745363	0.745454	0.745356	0.745284	0.745202	0.745219	0.745228	0.745285	0.745127	0.745194	0.74531	0.745196	0.745084
Hf178Hf1	77m_Int2	SE I	0.00028	0.00024	0.00024	0.00037	0.00024	0.00028	0.00021	0.0003	0.0003	0.00028	0.00024	0.00027	0.00026	0.00015	0.00032	0.00017	0.00015	0.00021	0.00021	0.00029	0.00031	0.00025	0.00021	0.00024	0.00031	0.00021	0.00022	0.00019	0.00023	0.0003	0.00026	0.0002	0.0002	0.00031	0.00022	0.00019	0.00021	0.00022	0.00017
-	Hf178Hf1	77m	1.48171	1.48096	1.48105	1.48108	1.48047	1.48036	1.4804	1.47997	1.4801	1.47997	1.47991	1.47996	1.47991	1.47966	1.47983	1.47991	1.48003	1.47994	1.47942	1.48203	1.48069	1.48097	1.48059	1.48065	1.48013	1.48018	1.48	1.48011	1.47992	1.48002	1.47974	1.47973	1.4797	1.47997	1.47967	1.47976	1.47976	1.47982	1.47944
AI176Hf1	77m_Int2	SE	0.0002	0.00019	0.00011	0.00021	0.00018	0.000086	0.00013	0.00013	0.00015	0.00017	0.00014	0.00013	0.00014	0.0001	0.00017	0.00012	0.00018	0.00021	0.00017	0.00022	0.00014	0.00013	0.00019	0.00012	0.00018	0.00019	0.00016	0.00017	0.00018	0.0002	0.00012	0.00016	0.00016	0.00018	0.00016	0.00015	0.000099	0.00018	0.00013
2	All176Hf1	77m	0.28975	0.28906	0.28992	0.28943	0.28961	0.288679	0.28658	0.29004	0.29016	0.28982	0.28982	0.29012	0.28666	0.28841	0.28968	0.28893	0.28835	0.29024	0.28956	0.28972	0.2898	0.28965	0.28962	0.28956	0.28969	0.28952	0.28957	0.2895	0.28957	0.2895	0.28936	0.28946	0.2894	0.28957	0.2894	0.28929	0.289253	0.28901	0.28911
Hf176Hf1	77mF_Int	2SE	0.000053	0.00005	0.000049	0.000061	0.000056	0.000056	0.000057	0.000057	0.000052	0.000051	0.000056	0.000046	0.000049	0.000065	0.000046	0.000055	0.000068	0.000063	0.000047	0.000048	0.000054	0.00004	0.000056	0.000032	0.000042	0.000052	0.00005	0.000049	0.00004	0.000047	0.000052	0.000055	0.000055	0.000053	0.000048	0.000057	0.000069	0.00005	0.000043
	Hf176Hf1	77mF	0.279547	0.279676	0.279686	0.279761	0.279739	0.279869	0.279736	0.279871	0.279891	0.279879	0.2799	0.279837	0.279859	0.279914	0.279932	0.279852	0.27989	0.279915	0.279919	0.279235	0.27943	0.279394	0.279447	0.279423	0.279565	0.279548	0.27958	0.279566	0.279597	0.279596	0.279585	0.279644	0.279586	0.279661	0.279683	0.279636	0.279641	0.279631	0.279628
Hf180Hf1	77c_Int2S	ш	0.00024	0.00028	0.00035	0.00022	0.00023	0.00017	0.00026	0.00021	0.0002	0.0003	0.0003	0.00024	0.00018	0.00028	0.00025	0.0002	0.00022	0.00026	0.00028	0.00021	0.00024	0.00025	0.00017	0.00021	0.00026	0.00022	0.00017	0.00016	0.00014	0.00017	0.00026	0.00022	0.00028	0.00018	0.00021	0.00022	0.00015	0.0002	0.00018
	Hf180Hf1	77c	1.88693	1.88692	1.88696	1.88661	1.88698	1.88685	1.88703	1.88724	1.88671	1.8868	1.88686	1.88688	1.8868	1.88689	1.88688	1.8868	1.88681	1.88674	1.88664	1.88681	1.88699	1.88693	1.88674	1.88699	1.88685	1.88701	1.88704	1.88677	1.8869	1.88685	1.88679	1.88694	1.8868	1.88694	1.88688	1.8869	1.88684	1.88691	1.88689
Hf178Hf1	77c_Int2S	ш	0.00034	0.00029	0.00027	0.00039	0.00026	0.00029	0.00026	0.0003	0.00037	0.00031	0.00025	0.00028	0.00026	0.00016	0.00033	0.00019	0.00016	0.00027	0.0002	0.00032	0.00038	0.0003	0.00025	0.00024	0.00033	0.00029	0.00024	0.00021	0.00025	0.00032	0.00029	0.00024	0.00025	0.00031	0.00024	0.00022	0.00017	0.00027	0.00019
	Hf178Hf1	77c	1.46713	1.46709	1.46711	1.46738	1.46708	1.4671	1.46696	1.46705	1.4671	1.46711	1.46702	1.46708	1.46705	1.46697	1.46706	1.46707	1.46719	1.46731	1.46691	1.46717	1.46688	1.46709	1.46701	1.46707	1.4671	1.46706	1.46715	1.46717	1.46708	1.46725	1.46704	1.46702	1.46699	1.46719	1.46706	1.46708	1.46696	1.46713	1.46686
Hf176Hf1	77cYb_Int	2SE	0.000096	0.00012	0.000074	0.00013	0.0001	0.0001	0.00011	0.0001	0.000089	0.00014	0.00012	0.00011	0.00012	0.000085	0.000097	0.00011	0.00015	0.00014	0.000091	0.00011	0.00011	0.000068	0.000082	0.000063	0.000066	0.00011	0.000099	0.0000	0.00008	0.000091	0.000099	0.000088	0.000089	0.000092	0.000094	0.0001	0.000081	0.00012	0.000077
	Hf176Hf1	77cYb	0.282392	0.28234	0.282334	0.28243	0.28236	0.2825	0.2824	0.28242	0.28236	0.28236	0.28249	0.28232	0.28245	0.282499	0.282416	0.28235	0.28243	0.28237	0.282319	0.28217	0.28216	0.282143	0.28212	0.282079	0.282102	0.2821	0.282116	0.282093	0.28211	0.282046	0.282012	0.282125	0.282042	0.28212	0.28222	0.28212	0.282113	0.28212	0.282061
Hf176Hf1	77cf_Int2	SE	0.000063	0.000059	0.000043	0.000067	0.000065	0.000059	0.000068	0.000061	0.000061	0.000063	0.000066	0.000059	0.00005	0.000066	0.000058	0.000063	0.000071	0.000068	0.00005	0.000059	0.000062	0.000052	0.00006	0.000041	0.000048	0.00006	0.000057	0.000047	0.000048	0.000052	0.00006	0.000067	0.000065	0.000056	0.00006	0.000066	0.000066	0.000055	0.00004
	Hf176Hf1	77cF	0.282345	0.28234	0.282344	0.282392	0.28231	0.282416	0.282317	0.282352	0.282388	0.28235	0.282376	0.282311	0.28233	0.282351	0.282385	0.282319	0.282355	0.282339	0.282323	0.282083	0.282079	0.282052	0.282054	0.282027	0.282064	0.282049	0.282046	0.28205	0.282062	0.282047	0.282021	0.282083	0.282027	0.282113	0.282106	0.282071	0.282097	0.282066	0.282041
	Comment	S	91500 - 1	91500 - 2	91500-3	91500 - 4	91500 - 5	91500-6	91500 - 7	91500 - 8	91500 - 9	91500 - 10	91500 - 11	91500 - 12	91500 - 13	91500 - 14	91500 - 16	91500 - 17	91500 - 18	91500 - 19	91500 - 20	GJ1 - 1	GJ1 - 2	GJ1-3	GJ1 - 4	GJ1 - 5	GJ1 - 6	GJ1 - 7	GJ1-8	GJ1-9	GJ1 - 10	GJ1 - 11	GJ1 - 12	GJ1 - 13	GJ1 - 14	GJ1 - 15	GJ1 - 16	GJ1 - 17	GJ1 - 18	GI1-19	GJ1-20

10000	_	Hf176Hf1		Hf176Hf1	-	Hf178Hf1		Hf180Hf1		Hf176Hf1		All176Hf1		Hf178Hf1		Hf179Hf1	
176	HH	77cF_Int2	Ht176Ht1	77cYb_Int H	H178H11	77c_Int25	Hf180Hf1	77c_Int25	Hf176Hf1	77mF_Int	AII176Hf1	77m_Int2	Hf178Hf1	77m_Int2	H1179Hf1	77m_Int2 +	11180H11
45		SE	77cYb	2SE .	77c E	ш	77c	ш	77mF	2SE	77m	SE	77m	SE	77m	SE	7m
G	82564	0.000076	0.2826	0.00011	1.46707	0.00041	1.88681	0.00026	0.27972	0.000058	0.280731	0.000053	1.48189	0.00037	0.74734	0.00011	1.94425
3	82537	0.000048	0.282558	0.000061	1.46719	0.00037	1.88681	0.0002	0.27984	0.00004	0.280855	0.000043	1.48123	0.00034	0.746555	0.000065	1.9412
	282571	0.000056	0.28259	0.000086	1.46705	0.0003	1.88682	0.0002	0.279872	0.00005	0.280762	0.000052	1.4811	0.00031	0.746571	860000.0	1.94128
1.1	282601	0.000055	0.28264	0.000076	1.46707	0.00029	1.88671	0.0002	0.27999	0.000048	0.281009	0.000046	1.48066	0.00027	0.7461	0.000093	1.93934
	282538	0.000031	0.282573	0.000067	1.46694	0.00019	1.88684	0.00022	0.280033	0.000036	0.281014	0.000035	1.48004	0.00019	0.745584	0.000069	1.93759
	282595	0.000054	0.282635	0.000073	1.46696	0.00026	1.88672	0.00014	0.280037	0.000042	0.281044	0.000039	1.48027	0.0002	0.745824	0.000097	1.93828
	282585	0.000046	0.282616	0.000065	1.46695	0.0003	1.88688	0.00021	0.280043	0.000038	0.281096	0.00004	1.48017	0.00026	0.745739	0.000001	1.93811
100	282574	0.000046	0.282641	0.000063	1.46708	0.00025	1.88701	0.00017	0.280084	0.00004	0.281174	0.00004	1.48003	0.00023	0.745461	0.000061	1.93716
	282519	0.000041	0.282503	0.000056	1.46708	0.00026	1.88682	0.00012	0.280073	0.000036	0.281153	0.00004	1.4798	0.00024	0.745236	0.000066	1.9361
_	.282576	0.000047	0.282598	0.000065	1.46711	0.00026	1.88676	0.00019	0.280118	0.000045	0.281081	0.000038	1.4799	0.00024	0.745297	0.000083	1.93627
-	1.282527	0.000048	0.28254	0.000069	1.46712	0.00013	1.88678	0.00015	0.280055	0.000043	0.281066	0.000045	1.47997	0.0001	0.745367	0.000074	1.93656
0	1.282546	0.000043	0.282574	0.000065	1.4671	0.00016	1.88694	0.00017	0.280104	0.000039	0.281139	0.00004	1.47981	0.00015	0.745215	0.000067	1.93613
0	.282548	0.00006	0.282547	0.000095	1.46707	0.00024	1.88708	0.00011	0.280107	0.000049	0.281105	0.000044	1.47977	0.0002	0.745206	0.00007	1.93624
0	.282594	0.00004	0.282653	0.00007	1.467	0.00022	1.88689	0.00017	0.280185	0.000033	0.281155	0.000032	1.47952	0.0002	0.745035	0.000062	1.93538
-00550	0.28262	0.000048	0.282638	0.000068	1.46697	0.00025	1.88677	0.00018	0.280158	0.000043	0.281238	0.000038	1.47978	0.00025	0.745315	0.000084	1.93635
0	.282575	0.000057	0.282584	0.000086	1.46695	0.00026	1.88692	0.00017	0.280039	0.000048	0.281101	0.00004	1.48014	0.00023	0.745707	0.000071	1.93802
0	.282523	0.000055	0.282536	0.000085	1.467	0.00017	1.88687	0.00017	0.280072	0.000046	0.281141	0.000049	1.47975	0.00016	0.745259	0.000065	1.93624
0	.282576	0.000044	0.282637	0.000057	1.46714	0.00018	1.88691	0.00018	0.280134	0.000035	0.281117	0.000041	1.47984	0.00019	0.74521	0.001	1.93611
1000	0.28256	0.000053	0.28261	0.000086	1.46706	0.00027	1.88695	0.00019	0.280045	0.000049	0.281053	0.000045	1.48015	0.00023	0.7456	0.00011	1.93764
0	1.282571	0.000033	0.28259	0.000073	1.46675	0.00024	1.88679	0.00015	0.280152	0.000028	0.281122	0.000036	1.47933	0.00024	0.745093	0.000074	1.93543
C	282524	0.000052	0 282538	0.000072	1 46699	0.00027	1 88687	0 00019	0 279673	0 000048	0 28565	0.00035	1 48184	0 00032	0 7473R	0.00016	1 94447
0 0	JOJEJC	100000		2.00000			1 00070	0,000,0	0.070005	0.0000	0 20500	110000	C101 1		CL345 0	10000	104100
1	0 28755	0.000058	0.28250	0.000072	1.46099 1.46715	0.00034	1.88693	0.00015	0.279906	0.000048	0.28538	0.00018	1 48091	0.00034	0.746787	1000000	1 94027
0	.282523	0.000041	0.282573	0.000066	1.467	0.00018	1.88694	0.00019	0.279906	0.000038	0.28541	0.00017	1.48062	0.00015	0.746137	0.000077	1.93972
0	.282545	0.000054	0.282571	0.000082	1.46708	0.00032	1.88686	0.00014	0.27994	0.000047	0.28509	0.00013	1.48063	0.00028	0.746071	0.000063	1.93938
0	1.282534	0.000045	0.282564	0.00007	1.46699	0.00019	1.88693	0.00022	0.280041	0.000042	0.28492	0.0002	1.47996	0.0002	0.745482	0.000053	1.93717
0	1.282541	0.000038	0.282536	0.000075	1.46696	0.00022	1.88672	0.00019	0.279963	0.000035	0.28478	0.00012	1.48037	0.00021	0.745928	0.000072	1.93868
0	1.282504	0.000052	0.282493	0.000064	1.46708	0.00021	1.88698	0.00017	0.279998	0.000046	0.284483	0.000086	1.48012	0.0002	0.745553	0.000067	1.93749
0	1.282499	0.00004	0.28251	0.000075	1.46702	0.00022	1.88682	0.00016	0.279992	0.000034	0.284742	0.000061	1.48007	0.00021	0.745559	0.000067	1.93735
0	1.282553	0.000046	0.282578	0.000066	1.46702	0.00025	1.88683	0.00015	0.280074	0.000037	0.2862	0.00012	1.47991	0.00022	0.745408	0.000064	1.93677
0	1.282529	0.000052	0.282561	0.000081	1.46714	0.00028	1.88708	0.00013	0.280084	0.000045	0.285838	0.000071	1.47986	0.00024	0.745232	0.000069	1.93635
0	1.282504	0.000059	0.28253	0.00011	1.467	0.00015	1.88692	0.00017	0.280052	0.000054	0.2875	0.00018	1.47976	0.00014	0.745269	0.00004	1.93632
0	1.282542	0.000057	0.28257	0.000068	1.46692	0.00024	1.88681	0.00016	0.280075	0.00005	0.28602	0.00012	1.47975	0.00022	0.745344	0.000062	1.93651
0	1.282507	0.000044	0.28253	0.000066	1.46701	0.00022	1.88693	0.00015	0.28007	0.000038	0.28636	0.00013	1.47968	0.00019	0.745188	0.000055	1.93602
0	1.282539	0.000032	0.282559	0.00006	1.46708	0.00028	1.8868	0.00014	0.280073	0.00003	0.2849	0.00041	1.4799	0.00025	0.745333	0.000063	1.93645
0	1.282506	0.000054	0.282544	0.000081	1.46703	0.00022	1.88682	0.00021	0.280078	0.000048	0.28838	0.00022	1.47965	0.00023	0.745136	0.00007	1.9357
-	1.282529	0.000042	0.282554	0.00007	1.46703	0.00014	1.88675	0.00013	0.280096	0.000036	0.28781	0.00015	1.47968	0.00014	0.745164	0.000093	1.93574
0	1.282526	0.000056	0.28254	0.000071	1.46712	0.00049	1.88679	0.00021	0.28008	0.000038	0.28806	0.00012	1.48016	0.00073	0.74515	0.00023	1.93572
-	1.282536	0.000043	0.282596	0.000049	1.46707	0.0002	1.88676	0.00022	0.280112	0.000035	0.28753	0.00022	1.47968	0.00017	0.745117	0.000082	1.93557

	If180Hf1	"Zm	1.93807	1.93716	1.93564	1.93547	1.93556	01000 1	1.93919	1.93664	1.93522	1.93487	1.93597	1.9392	1.93679	1.93576	1.93571	1.93491	1.94363	1.9412	1.93952	1.93851	1.93836	1.93878	1.9372	1.93739	1.93597	1.93682	1.9369	1.93631	1.93566	1.93606	1.93561	1.93661	1.93589	1.93736	1.93458	1.93515
Hf179Hf1	77m_Int2_H		0.000078	0.000068	0.000055	0.000078	0.000068	0.00000	520000.0	0.000045	0.000062	0.000068	0.00012	0.000063	0.000052	0.00006	0.00009	0.000083	0.000094	0.0001	0.000081	0.000095	0.000073	0.00007	0.000077	0.000087	0.000068	0.000093	0.000092	0.000078	0.000049	0.000074	0.000079	0.0001	0.000072	0.00011	0.000074	0.000065
-	H179Hf1	7m	0.745729	0.745509	0.745131	0.745076	0.745085	000772.0	0./46002	0.74533	0.745026	0.744933	0.74523	0.746036	0.745406	0.745136	0.745167	0.74492	0.747163	0.74654	0.74609	0.745799	0.745772	0.745853	0.74549	0.745534	0.745135	0.745394	0.745457	0.745234	0.74505	0.745208	0.745074	0.74534	0.745173	0.74554	0.744804	0.745
Hf178Hf1	77m_Int2_1	13	0.00018	0.00019	0.00015	0.00023	0.00014	00000	0.0003	0.00015	0.00022	0.0002	0.00021	0.00022	0.0002	0.00023	0.00021	0.00014	0.00033	0.00026	0.00025	0.0002	0.00029	0.00024	0.00019	0.00017	0.00027	0.00018	0.00019	0.00019	0.00018	0.00019	0.00025	0.00014	0.00021	0.00026	0.00021	0.00021
	Hf178Hf1	77m	1.4804	1.48008	1.47977	1.47952	1.47971	1001 1	L.48U5	1.47989	1.47953	1.47943	1.4797	1.48059	1.47996	1.47971	1.47963	1.47955	1.48177	1.48113	1.48067	1.48034	1.48036	1.48037	1.48001	1.48012	1.47975	1.47994	1.48016	1.47977	1.47958	1.47972	1.47967	1.47988	1.47978	1.4802	1.47942	1.47952
AII176Hf1	77m_Int2_1		0.00031	0.0012	0.00067	0.0004	0.0028		7700.0	0.0012	0.0021	0.0014	0.0012	0.0038	0.0038	0.0038	0.0052	0.0041	0.00072	0.0043	0.00054	0.0013	0.00089	0.00012	0.00096	0.0085	0.007	0.0034	0.0014	0.0028	0.0012	0.01	0.01	0.0051	0.0035	0.0021	0.0054	0.0011
	VII176Hf1 7	7m 3	0.33869	0.332	0.29589	0.3313	0.3074	1000	0.4335	0.4	0.4914	0.4032	0.3954	0.5587	0.5653	0.5516	0.5542	0.5618	0.31467	0.3764	0.36325	0.3202	0.33693	0.32718	0.3573	0.4118	0.3918	0.3266	0.3126	0.3574	0.3843	0.438	0.402	0.3468	0.4052	0.3368	0.4711	0.3181
Hf176Hf1	77mF_Int /	SE	0.000034	0.000032	0.000036	0.000037	0.000036	F# 0000 0	U.UUUU4/	0.000049	0.000053	0.000044	0.000034	0.000052	0.000047	0.000037	0.000051	0.000052	0.000055	0.000045	0.000037	0.000035	0.000056	0.000036	0.000038	0.000049	0.000048	0.00004	0.000041	0.000034	0.000052	0.000051	0.000038	0.000037	0.00004	0.000054	0.000058	0.000033
	H176Hf1	77mF	0.279623	0.279689	0.27974	0.279732	0.279731	O POPPEO	9556/7.0	0.279719	0.279773	0.279779	0.279691	0.279512	0.279607	0.279668	0.279664	0.279653	0.28	0.280018	0.280182	0.280188	0.280214	0.280218	0.280268	0.28026	0.28032	0.280297	0.2803	0.280303	0.280352	0.28034	0.280392	0.280319	0.280307	0.280216	0.280334	0.280355
Hf180Hf1	77c_Int25	ш	0.0002	0.00019	0.00015	0.00021	0.00019	110000	TTOOO'O	0.00018	0.00019	0.00021	0.00015	0.00014	0.00015	0.00019	0.00021	0.00019	0.00021	0.00026	0.00021	0.00021	0.00016	0.00017	0.00016	0.00019	0.00019	0.00024	0.00025	0.00018	0.00019	0.00015	0.00017	0.00018	0.00021	0.00018	0.0002	0.00021
	Hf180Hf1	77c	1.88688	1.88683	1.88678	1.88682	1.88687	1 00001	L.88094	1.88692	1.88676	1.88677	1.8867	1.88682	1.88685	1.88688	1.88671	1.88686	1.88686	1.88686	1.88693	1.88705	1.887	1.8871	1.88694	1.88695	1.88708	1.88693	1.88677	1.88704	1.8871	1.88689	1.88694	1.88694	1.88686	1.88691	1.88698	1.88679
Hf178Hf1	77c_Int2S		0.00021	0.00021	0.00017	0.00024	0.00015	100000	0.00034	0.00014	0.00024	0.00022	0.0002	0.00025	0.00021	0.00025	0.00025	0.00017	0.00039	0.0003	0.00028	0.00021	0.00033	0.00026	0.00027	0.00018	0.0003	0.0002	0.00023	0.00023	0.00019	0.00022	0.00027	0.00017	0.00022	0.00027	0.00022	0.00021
	Hf178Hf1	77c	1.46719	1.46708	1.46715	1.46696	1.46713	COLUMN F	1.46/UZ	1.46705	1.46701	1.46701	1.46698	1.46707	1.46707	1.46709	1.46697	1.46714	1.46713	1.46711	1.4671	1.46706	1.4671	1.46703	1.4671	1.4671	1.46713	1.46706	1.46723	1.46705	1.46704	1.46703	1.46711	1.46706	1.46711	1.46717	1.46712	1.46703
Hf176Hf1	77cYb_Int	2SE	0.000078	0.000064	0.000074	0.000069	0.000075	C 1000 0	ZTOUD.D	0.000092	0.00014	0.00008	0.00007	0.00012	0.00011	0.000079	0.001	0.00013	0.0001	0.00011	0.000095	0.000073	0.00011	0.000067	0.000082	0.000079	0.000089	0.000087	0.000077	0.000079	0.0001	0.00014	0.000098	0.000089	0.000091	0.0001	0.00017	0.000076
	Hf176Hf1	77cYb	0.282185	0.282242	0.282182	0.282226	0.282137	000000	0.28229	0.282313	0.28227	0.282301	0.282271	0.28228	0.28234	0.28234	0.28219	0.28227	0.28291	0.28276	0.282893	0.282787	0.28287	0.282902	0.282883	0.282897	0.282861	0.282851	0.282869	0.282809	0.28285	0.28296	0.282881	0.2829	0.282806	0.28275	0.28284	0.282837
Hf176Hf1	77cf_Int2	SE	0.000041	0.000036	0.000041	0.000038	0.000035	0.000011	<<000000	0.000047	0.000052	0.000041	0.000033	0.000053	0.000047	0.000038	0.000047	0.000054	0.000066	0.000052	0.000039	0.000037	0.000066	0.000044	0.000043	0.000049	0.000048	0.000044	0.000049	0.00004	0.000054	0.000051	0.00005	0.00005	0.000037	0.000063	0.00006	0.00004
	Hf176Hf1	77cF	0.282159	0.282184	0.282164	0.282145	0.282145		0.282143	0.282184	0.282177	0.282165	0.282133	0.282106	0.282082	0.282092	0.282093	0.282035	0.282814	0.282713	0.282792	0.282743	0.282765	0.282784	0.282765	0.282765	0.282749	0.282776	0.28279	0.28275	0.282765	0.282784	0.282803	0.282787	0.282742	0.282721	0.2827	0.282759
	Comment	s	MUN 1 - 1	MUN 1 - 2	MUN 1 - 3	MUN 1 - 4	MUN 1 - 5			MUN 3 - 2	MUN 3 - 3	MUN 3 - 4	MUN 3 - 5	MUN 4 - 1	MUN4-2	MUN 4 - 3	MUN 4 - 4	MUN 4 - 5	R33 - 1	R33 - 2	R33 - 3	R33 - 4	R33 - 5	R33 - 6	R33 - 7	R33 - 8	R33 - 9	R33 - 10	R33 - 11	R33 - 12	R33 - 13	R33 - 14	R33 - 15	R33 - 16	R33 - 17	R33 - 18	R33 - 19	R33 - 20

TE-06 0.282294 0.000063 BE-07 0.282295 0.000059 BE-07 0.282295 0.000043 BE-07 0.282295 0.000067 BE-07 0.282295 0.000067 BE-07 0.282295 0.000067 BE-07 0.282244 0.000067 BE-07 0.282243 0.000065 BE-06 0.282271 0.000068
1.7E-06 0.282294 0.0000 8.8E-07 0.28229 0.0000 2.3E-06 0.282295 0.0000 9.3E-07 0.282344 0.0000 9.2E-07 0.282354 0.0000 3.3E-06 0.28237 0.0000 1.1E-06 0.282271 0.0000
8.8E-07 0.28229 (2.3E-06 0.282295 (9.3E-07 0.282344 (9.2E-07 0.282263 (3.3E-06 0.282263 (1.1E-06 0.282271
311 2.3E-06 0.282 311 9.3E-07 0.282 311 9.2E-07 0.282 312 3.3E-06 0.283 311 9.2E-07 0.282 311 9.2E-06 0.283 311 9.2E-06 0.283 311 1.1E-06 0.283
000311 2.3E-06 000311 9.3E-07 0.000311 9.2E-07 0.000312 3.3E-06 0.000312 3.3E-06
0.2 0.000311 0.29 0.000311 0.24 0.000312 0.27 0.000311
4.14 0.29 4.23 0.24 4.23 0.24 4.01 0.27
-06 4.23 -06 4.23 -07 4.01
1249 2.6E-06 198 7.2E-07
0.00012
0.01025 0.01035 0.01001
0.00024 0.01025 0.00037 0.01035 0.00028 0.01001

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	Age																																							
StdCorr_1 76Hf 177	Hf_Int2SE	0.000076	0.000048	0.000056	0.000055	0.000031	0.000054	0.000046	0.000046	0.000041	0.000047	0.000048	0.000043	0.00006	0.00004	0.000048	0.000057	0.000055	0.000044	0.000053	0.000033	0.000052	1000000	0.000050	\$CUUUUU.U	0.000014	5000000	0.000038	0.000052	0.00004	0.000046	0.000052	0.000059	0.000057	0.000044	0.000032	0.000054	0.000042	0.000056	0.000043
StdCorr_1 76Hf 177	Hf	0.282513	0.282486	0.282522	0.282553	0.282491	0.282549	0.28254	0.282529	0.282475	0.282534	0.282485	0.282505	0.282508	0.282554	0.282582	0.282538	0.282486	0.282541	0.282526	0.282538	27773C D		0.282485	106282.0	C14282.0	164202.0	0.282495	0.28246	0.282456	0.28251	0.282487	0.282463	0.282501	0.282468	0.282502	0.282469	0.282493	0.282491	0.282503
StdCorr_1	Hf_Int2SE	4.5E-07	4.3E-07	4.4E-07	4.5E-07	4E-07	4.5E-07	5.5E-07	5.6E-07	4.6E-07	6.3E-07	4.2E-07	4.3E-07	6.7E-07	6.6E-07	4.8E-07	4.1E-07	8.4E-07	4E-07	4.6E-07	4.3E-07	6 6E-D6	100 100	4.4E-UD	4.8E-U0	5.4E-06	6.6E-06	4.9F-06	1.1E-06	1.8E-06	3.7E-06	1.4E-06	1.5E-06	1.1E-06	9.6E-07	9.3E-06	0.000002	0.000001	1.5E-06	2.4E-06
StdCorr_1 76Lu 177	Hf	2.6E-05	2.7E-05	2.2E-05	2.67E-05	2.53E-05	2.91E-05	3.71E-05	2.7E-05	2.68E-05	0.000025	2.55E-05	2.56E-05	3.72E-05	2.86E-05	2.76E-05	2.83E-05	2.93E-05	2.88E-05	2.54E-05	2.58E-05	0.00013	000000	0.0000	21000.0	0.0001123	0.000119	0.000143	9.29E-05	0.000101	0.000131	0.000118	0.000164	0.00019	0.000156	0.000107	0.0002	0.000193	0.000162	0.000164
alHf In	l u	0.29	0.38	0.45	0.49	0.45	0.45	0.37	0.44	0.4	0.43	0.44	0.41	0.42	0.4	0.42	0.39	0.43	0.37	0.43	0.44	0 G		10.0	0.47	0.03	0.48	0.46	0.51	0.48	0.46	0.45	0.39	0.48	0.48	0.43	0.41	0.47	0.44	0.33
Tot	FotalHf t25I	2.97	6.76	6.88	6.99	6.97	7.1	7.12	7.1	7.01	7.2	7.06	6.97	7.14	6.93	6.73	6.46	6.69	6.71	6.8	6.88	6 92		14.1	15.1	6/./	P.04	7.83	7.85	7.85	7.69	7.87	7.21	7.45	7.43	7.57	7.43	7.56	8.19	7.77
1176Hf1	77_Int2SE	4E-07	3.6E-07	4E-07	3.8E-07	3.5E-07	3.6E-07	3.6E-07	5.2E-07	4.2E-07	5.6E-07	3.9E-07	3.9E-07	4.2E-07	5.2E-07	4.3E-07	3.6E-07	6.9E-07	3.2E-07	4.3E-07	3.8E-07	5 RE-DR		3.7E-UD	4.4E-UD	4.6E-Ub	5.1E-06	3.2F-06	0.000001	1.6E-06	3.2E-06	1.3E-06	1.4E-06	6.9E-07	7.5E-07	8.2E-06	1.6E-06	7.8E-07	1.4E-06	2.1E-06
Lu176Hf1	11	2.29E-05	2.26E-05	2E-05	2.28E-05	2.24E-05	2.3E-05	2.38E-05	2.5E-05	2.44E-05	2.22E-05	2.33E-05	2.36E-05	2.33E-05	2.28E-05	2.5E-05	2.49E-05	2.42E-05	2.26E-05	2.34E-05	2.25E-05	0.000114	0.000115	0.000100	0.000109	COLUUU.U	CU-300.6	9.27E-05	8.64E-05	9.13E-05	0.000116	0.000109	0.000149	0.000118	0.000125	0.000094	0.000164	0.000151	0.000149	0.000143
vb176Hf1	77_Int2SE	0.000021	0.00002	0.00002	0.000019	0.000019	0.000023	0.000018	0.000024	0.000017	0.000021	0.000018	0.000021	0.000021	0.000022	0.000021	0.000023	0.00002	0.000021	0.000024	0.000024	0 00035	0,0000	0.00016	910000	/1000.0	21000.0	0.00011	0.000074	0.000047	0.00011	0.000072	0.00017	0.000092	0.00015	0.00041	0.00021	0.00016	0.00015	0.00021
vh176Hf1	11	0.00103	0.001031	0.000905	0.001034	0.000994	0.001021	0.001068	0.001104	0.001094	0.000975	0.001023	0.001048	0.00101	0.000981	0.001093	0.001075	0.001083	0.00095	0.001022	0.000981	0.00611	110000	0.00500	0.0013	0.0056	C2C00.0	0.0049	0.004564	0.00483	0.00623	0.005848	0.00756	0.006037	0.00638	0.00491	0.00842	0.00784	0.00814	0.00753
Hf180Hf1 77m Int2	SE	0.00035	0.00028	0.00029	0.00034	0.00027	0.00029	0.00032	0.00024	0.00022	0.00022	0.00027	0.00021	0.00023	0.00015	0.00029	0.00024	0.00021	0.00031	0.0005	0.00019	0 00066	00000	0.0000	6700000	0.00022	0,0002	0.00071	0.00019	0.00024	0.0002	0.00027	0.00018	0.00022	0.00015	0.00019	0.00023	0.00035	0.00096	0.00019
Comment	s	MUD - 1	MUD-2	MUD - 3	MUD - 4	MUD - 5	MUD - 6	MUD - 7	MUD - 8	MUD - 9	MUD - 10	MUD - 11	MUD - 12	MUD - 13	MUD - 14	MUD - 15	MUD - 16	MUD - 17	MUD - 18	MUD - 19	MUD - 20	DI EC. 1		PLES- Z	PLES-3	PLES-4	PIFS-6	PLFS - 7	PLES - 8	PLES - 9	PLES - 10	PLES - 11	PLES - 12	PLES - 13	PLES - 14	PLES - 16	PLES - 17	PLES - 18	PLES - 19	PLES - 20

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-		Age		- C		00	Б	10	2	2	1	3	~	1	80	2	4	1	0	2	6	1	5	4	ŝ	6	80	4	6	4	4	Ч	5	5	1		9	4
StdCorr_1	//T_1HQ/	Hf_Int2SE	0.00004		*n0000	0.00003	0.00003	0.00005	0.00004	0.00005	0.00004	0.00003	0.00005	0.00004	0.00003	0.00004	0.00005		0.00000	0.00005.	0.00003	0.00003	0.00006	0.00004	0.00004	0.00004	0.00004	0.00004	0.00004	0.0000	0.000054	0.00005	0.0000	0.0000	0.00003	0.00006.	0.0000	0.0000.
StdCorr_1	//1-11/0/	Hf	0.282111	CC1207.0	771707.0	0.282106	0.28211	0.282096	0.282139	0.282135	0.282126	0.282098	0.282058	0.282037	0.28205	0.282055	0.282		U.282/03	0.282663	0.282743	0.282695	0.282717	0.282737	0.282719	0.28272	0.282705	0.282733	0.282748	0.28271	0.282725	0.282745	0.282764	0.282749	0.282706	0.282685	0.282666	0.282726
StdCorr_1	/0rn_1//	Hf_Int2SE	0.000024 2 0F 0F	0.55.00	00-30.2	0.000015	0.000069	6.1E-06	9.5E-06	0.000022	8.2E-06	8.1E-06	0.000012	0.000025	0.000016	0.00004	0.000039		5./E-Ub	0.000093	0.000025	0.000024	0.000011	0.000022	0.000023	0.00017	0.00014	0.00013	0.000019	0.000044	0.000061	0.00025	0.00018	0.00011	0.00017	0.000049	0.000074	0.000011
StdCorr_1	//1-110/	Ħ	0.001631	0.000466	0.000400	0.001408	0.000905	0.004532	0.003608	0.005469	0.003453	0.003606	0.007945	0.007926	0.007059	0.007632	0.008616		//nTnn'n	0.002908	0.002329	0.001303	0.001725	0.001645	0.00317	0.00352	0.00309	0.00143	0.000926	0.002237	0.004503	0.00504	0.00336	0.002	0.0039	0.001966	0.005206	0.001213
al fillet	DIAIHT_IN	SE	0.39	110	0.41	0.41	0.33	0.4	0.42	0.43	0.49	0.47	0.5	0.47	0.48	0.45	0.45		0.34	0.32	0.37	0.41	0.42	0.43	0.41	0.29	0.33	0.4	0.36	0.35	0.37	0.35	0.28	0.29	0.32	0.29	0.31	0.36
÷	2	FotalHf t2	7 21	10-1	10.0	7.27	7.6	7.19	7.26	7.15	8.18	8.08	8.98	9.22	9.17	8.76	8.47		TA'C	5.95	6.45	6.99	7.09	7.02	6.74	5.9	6.16	6.2	6.7	6.02	6.24	5.88	6.14	6.08	5.49	6.27	5.07	6.23
131132111	T1HQ/TN1	77_Int2SE	0.000021 2 1F AF	0,00000	5000000	0.000014	0.000055	5.3E-06	9.1E-06	0.000022	7.5E-06	6.9E-06	0.00001	0.000024	0.000016	0.000036	0.000032	100000 0	suuuuu.u	0.000078	0.000022	0.000021	90-39.6	0.000016	0.000015	0.00016	0.00013	0.00012	0.000018	0.00004	0.000038	0.0002	0.00017	0.000096	0.00014	0.000038	0.000069	90-36-06
640.36 6.0	TIH9/TN1	77	0.001404	TOTTOO O	0.000442	0.00126	0.000728	0.00391	0.003164	0.005213	0.003097	0.002915	0.006866	0.007017	0.006748	0.006857	0.007		0.000944	0.002448	0.002114	0.001111	0.00153	0.001265	0.002068	0.00329	0.0028	0.00127	0.000855	0.002	0.0028	0.00408	0.00304	0.001747	0.0032	0.00155	0.004839	0.001051
671132.649	TIHQ/TQ	7_Int2SE	0.00033	000000	600000	0.00042	0.0029	0.0023	0.0012	0.0022	0.0015	0.0013	0.004	0.004	0.004	0.0053	0.0042	100000	c/nnn.n	0.0044	0.00056	0.0013	0.00094	0.00012	0.001	0.0086	0.0071	0.0034	0.0014	0.0029	0.0012	0.01	0.011	0.0052	0.0035	0.0022	0.0056	0.0011
V 19110614	L TIHO/TO	1 11	0.05981	202100	070TA'A	0.0521	0.0279	0.1558	0.1214	0.2137	0.1245	0.1169	0.2827	0.2888	0.2746	0.2772	0.2848		710000	0.0977	0.08405	0.0404	0.05725	0.04741	0.0777	0.1329	0.1126	0.0466	0.0326	0.0778	0.1048	0.159	0.122	0.0671	0.126	0.0571	0.1924	0.038
Hf180Hf1		SE	0.00026		etonn'n	0.00023	0.0002	0.00032	0.00016	0.0002	0.00024	0.00042	0.00026	0.00019	0.00024	0.00026	0.00028		0.00033	0.0003	0.00025	0.00031	0.00024	0.00026	0.00025	0.00028	0.00027	0.00031	0.00026	0.00024	0.00015	0.0002	0.00029	0.00034	0.00024	0.00035	0.00025	0.00021
+	Comment	S	MUN 1-1		C-T NOM	MUN 1-4	MUN 1 - 5	MUN 3 - 1	MUN 3 - 2	MUN 3 - 3	MUN 3 - 4	MUN 3 - 5	MUN 4 - 1	MUN 4 - 2	MUN 4 - 3	MUN 4 - 4	MUN 4 - 5		K33 - I	R33 - 2	R33 - 3	R33 - 4	R33 - 5	R33 - 6	R33 - 7	R33 - 8	R33 - 9	R33 - 10	R33 - 11	R33 - 12	R33 - 13	R33 - 14	R33 - 15	R33 - 16	R33 - 17	R33 - 18	R33 - 19	R33 - 20

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Appendix F - Fig F1: Comparison of measured Hf standards relative to accepted values. Analyses of reference zircons (91500, GJ-1, Mudtank, Plešovice, R33 and MUN 1,3,4) were conducted throughout the session to verify accuracy and ensure reproducibility. These ratios are within the range of TIMS derived values reported for 91500 by Wiedenbeck, et al. (1995) (0.282290 \pm 0.000014), and LA-ICP-MS values for GJ-1 (0.282000 \pm

0.000005) (Morel, et al., 2008), Mudtank (0.282504 ± 0.000044) (Woodhead & Hergt, 2005), Plešovice (0.282482 ± 0.000013) (Sláma, et al., 2008), R33 (0.282764 ± 0.000014) (Fisher, et al., 2014) and doped synthetic zircon standards MUN 1, MUN3 and MUN 4 (0.282140 ± 0.000006 , 0.282135 ± 0.000005 , 0.282141 ± 0.000006 , respectively) (Fisher, et al., 2011).

References

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