

**GEOLOGY, MINERALOGY, GEOCHEMISTRY, AND
ALTERATION PATTERNS OF THE BANANA ZONE COPPER-
SILVER MINERALISATION, GHANZI-CHOBE BELT,
NORTHWEST BOTSWANA**

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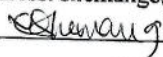
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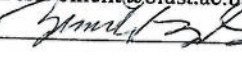
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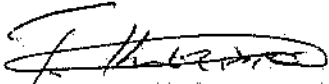
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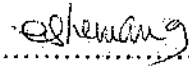
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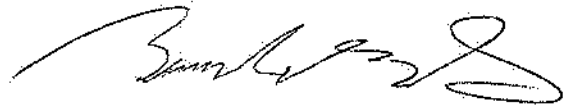
CERTIFICATION

The undersigned certifies that he has read and hereby recommends for acceptance by the College of Science a thesis titled: **GEOLOGY, MINERALOGY, GEOCHEMISTRY, AND ALTERATION PATTERNS OF THE BANANA ZONE COPPER-SILVER MINERALISATION, GHANZI-CHOBE BELT, NORTHWEST BOTSWANA**, in fulfilment of the requirements for the degree of Master of Science in Geology of the BIUST.



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ABSTRACT

The Banana Zone Cu-Ag Prospect is within the SW-NE trending Ghanzi-Chobe Belt (GCB), which extends from the Namibia-Botswana border in the south to Shinamba Hills in the north near the Botswana-Zambia border. The Banana Zone is also part of the Ghanzi Project owned by Khoemacau Copper Mining and situated between the Ghanzi and Maun townships, in Botswana. The Banana Zone has inferred resource of over 225.4 million metric tonnes (Mt) grading 0.64% Cu and 8.1 g/t Ag. The GCB consists of deformed and metamorphosed (greenschist) metasedimentary rocks of the Ghanzi Group, that is commonly subdivided into four formations: the Kuke Formation, Ngwako Pan Formation, D'Kar Formation, and Mamuno Formation, in ascending stratigraphic order. In this study only two formations including the Ngwako Pan and the D'Kar Formations were investigated. The Ngwako Pan Formation is made up of continental siliciclastic sediments, mainly immature to submature sandstone with siltstone and shale intercalations, while the D'Kar Formation contains shallow marine to lacustrine grey-green and locally carbonaceous mudrocks (siltstone and mudstone/shale) with fine-grained sandstone and minor limestone to marlstone. The Cu-Ag mineralisation is hosted at the contact between hematite-rich Ngwako Pan and reduced organic matter-bearing D'Kar Formations. The presence of reduced organic matter is likely to have controlled Cu-Ag precipitation by providing a reducing environment required for the precipitation of sulphides.

Petrographic and geochemical data were combined in order to decipher the petrogenesis of the Neoproterozoic clastic sedimentary succession of the Ngwako Pan and the D'Kar Formations associated with the Banana Zone Cu-Ag mineralisation. Sandstone of the Ngwako Pan and the D'Kar Formations is arkose and subarkose, composed of quartz (Q), feldspars (F), and lithic fragments (L). Moreover, geochemically the sandstone of the Ngwako Pan and the D'Kar Formations is considered as potassic and classified as arkose. Therefore, sandstone of both the Ngwako Pan and the D'Kar Formations is chemically and compositionally immature. On the other hand, mudrocks of the D'Kar Formation are finely laminated and are dominated by muscovite, sericite, chlorite, and quartz. The carbonate rocks of the D'Kar Formation are composed of pinkish-white crystalline limestone and yellowish grey or greyish black marlstone characterised by laminations that

are marked by undulating marly and silty layers that give the rock a “zebra-skin” appearance. The marlstone exhibits variable degree of deformation from shearing to folding and recrystallisation.

The modified chemical index of weathering (CIW') values indicated an intense chemical weathering of the source rock. The dominance of detrital quartz and feldspar grains coupled with $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios (average 29.67 and 24.52 for the sandstone of the Ngwako Pan and D'Kar Formations, respectively) and Ni and Cr depletion in this sandstone, suggests a dominant felsic source. However, high concentrations of Ni and Cr and a low $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio (<20) in the mudrocks of the D'Kar Formation indicate a mixed source. Provenance of the investigated sandstone and mudrocks samples is further supported by the REE patterns, the size of Eu anomaly, as well as La/Co, Th/Co, Th/Cr, and Cr/Th ratios, which show a felsic source for the sandstone of both the Ngwako Pan and D'Kar formations and an intermediate source for the mudrocks of the D'Kar Formation. Detrital modes (QFL diagrams) and geochemical characteristics of the sandstone of the Ngwako Pan and D'Kar Formations indicate that the detritus were probably supplied from a heavily weathered felsic continental block and deposited in a continental rift setting (passive margin) in a humid environment. The source rocks might have been the Palaeoproterozoic basement rocks (granitoids and granitic gneiss) and the Mesoproterozoic Kgwebe volcanic rocks exposed north of the study area. In addition, these basement rocks underlying the Neoproterozoic sedimentary succession that host the mineralisation, demonstrate the largest potential as metal source.

Numerous mineralisation styles were identified within the Banana Zone. These include: (i) disseminated mineralisation; (ii) vein-controlled mineralisation; (iii) breccia-related mineralisation; (iv) cleavage-controlled mineralisation; and (v) micro-pore space filling mineralisation styles. The hypogene ore phases are dominated by chalcopyrite, bornite, and chalcocite. Sphalerite and galena occur in minor quantities and mostly observed in veins. On the other hand, the supergene ore phases are dominated not only by chalcocite, digenite, and covellite, all formed at the expenses of chalcopyrite, bornite, and chalcocite, but also malachite. As no Ag-bearing mineral was observed, Ag is possibly occurring in chalcocite and in tetrahedrite/tennantite and galena as isomorphic admixtures.

Sulphides display complex textural relationships that demonstrate multiple mineralising events spanning from diagenesis through metamorphism and deformation to post-tectonic events. The early diagenetic to late diagenetic mineralisation is supported by the presence of framboidal pyrite, nonfracture-controlled distribution of both sulphides and gangue phases and replacement textures of Cu-sulphides after diagenetic cement and pyrite. Furthermore, the occurrence of sulphides in pre-folding bedding-parallel veins and brecciated zones supports early tectonic mineralisation, while the mineralisation in veins that did not experience deformation demonstrates a post-tectonic stage.

The host rocks are variably hydrothermally-altered and alteration occurs as pervasive and vein-controlled signifying a complex fluid flow history. The typically observed evolution is from high-temperature sodic-calcic (albitisation) to K-silicates (phyllic) to low temperature Mg-rich (propylitic) and carbonate-quartz assemblages. Most of the Cu-sulphides mineralisation occurs predominantly within zones of Mg-rich (propylitic) and carbonate-quartz alteration. Supergene processes are also observed in areas of high permeability, and have played a vital role in the remobilisation and upgrading of the primary mineralisation.

The temporal relationship between hydrothermal alteration assemblages and hypogene ore minerals indicates that the mineralisation took place at a temperature range of $\leq 60 - 400^{\circ}\text{C}$ and under weakly acidic to near-neutral and alkaline pH conditions. The mineralising fluids possibly were moderately saline (as supported by fluid inclusion data from a similar study at Boseto Mine (15-20 wt.% equiv. NaCl)), dense and moderately oxidised during earlier phases and became highly saline during the tectonic period. On the other hand, the association of chalcocite with iron oxide/hydroxide suggests a second generation of supergene chalcocite, which forms at temperature below 103°C and pH of about 3.5, whereas malachite developed at near-neutral to alkaline pH (4-9) conditions.

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CHAPTER 1

GENERAL INTRODUCTION

1.1. Background

The Banana Zone Cu-Ag deposit (hereafter referred to as Banana Zone, Fig.1.1), is within the SW-NE trending Kalahari Copper Belt (KCB) (Fig. 1.1a) that forms the southwestern annex of the Central African Copper Belt (CAC) in Botswana (Ghanzi-Chobe Belt) and in Namibia (Damara Belt). The Banana Zone is also part of the world's most extensive metallogenic province of sediment-hosted copper deposits (SHC) (**Sillitoe et al., 2010**). The Banana Zone (by analogy to its banana-like shape, see Fig. 1.1b) is an exploration site in the Ghanzi Project area owned by Khoemacau Copper Mining and consists of a simple doubly plunging folded structure with no major faulting (**GeoLogix, 2009**). The Ghanzi Project is composed of numerous mineralised zones, including three main zones from south to north the Banana Zone, Zone 5, and Boseto Mine (Fig. 1.1b). The Banana Zone, which is the focus of this study, has inferred resource worth over 225.4 million metric tonnes (Mt) grading 0.64 % Cu and 8.1 g/t Ag (**Hana Mining Limited, 2011**).

The Neoproterozoic metasedimentary succession hosting the Banana Zone forms part of the Ghanzi Group that was deposited in a rift basin (**Modie, 1996b; Kelepile et al., 2017**). This Group is commonly subdivided into four formations: the Kuke, Ngwako Pan, D'Kar, and Mamuno Formations, in ascending stratigraphic order (**Modie et al., 1998**). However, this study only focuses on two formations namely; the Ngwako Pan and the D'Kar Formations. The Ngwako Pan Formation is made up of continental siliciclastic sediments, mainly immature to submature hematite-rich sandstone with siltstone and shale intercalations, while the D'Kar Formation contains shallow marine to lacustrine grey-green and locally carbonaceous, pyrite-bearing mudrocks (siltstone and mudstone) with fine-grained sandstone and minor limestone to marlstone. The mineralisation occupies a stratigraphic redox boundary between the continental redbeds of the Ngwako Pan Formation and the grey-green transgressive marine facies of the D'Kar Formation (www.discoverymetals.com.au/dml_projects/boseto_copper_project). This stratabound mineralisation consists of Cu-Ag sulphide associated with sulphides of Fe, Zn, and Pb. The main ore minerals in decreasing order of abundance are chalcocite, bornite,

chalcopyrite, pyrite, sphalerite, and galena ([www.discoverymetals.com.au/dml projects/boseto copper project](http://www.discoverymetals.com.au/dml/projects/boseto_copper_project)). Non-Cu sulphides minerals of secondary origin that include azurite, malachite, and chrysocolla are also present, but in low concentrations. It is important to indicate that the mineral paragenesis and the source origin of the mineralisation in this area are not well understood. **Modie (2000)** showed that the deposits occur as disseminated mineralisation of inter-granular clusters or aggregates distributed through porous sandstone and as fracture-related mineralisation associated with secondary porosity.

Economic interest in the Ghanzi-Chobe Belt (GCB) dates back to the early 1960s with several exploration groups (Table 1.1) working in the area (e.g., **Carney et al., 1994**). However, the first in-depth study of the GCB was conducted by **Schwartz et al. (1995)** in the Ngwako Pan copper deposits (now known as Boseto). These investigations resulted in the discovery of mineralised strata of the Ghanzi Group in the Ngwako Pan area. During the recent years, exploration activities in the area intensified, resulting in the successful identification of significant copper-silver resources (**Botswana Resource Sector Report, 2013**). This breakthrough resulted in the opening of Boseto Mine by Discovery metals in 2011, the first mine in the Botswana section of the belt, that later closed down in 2014.

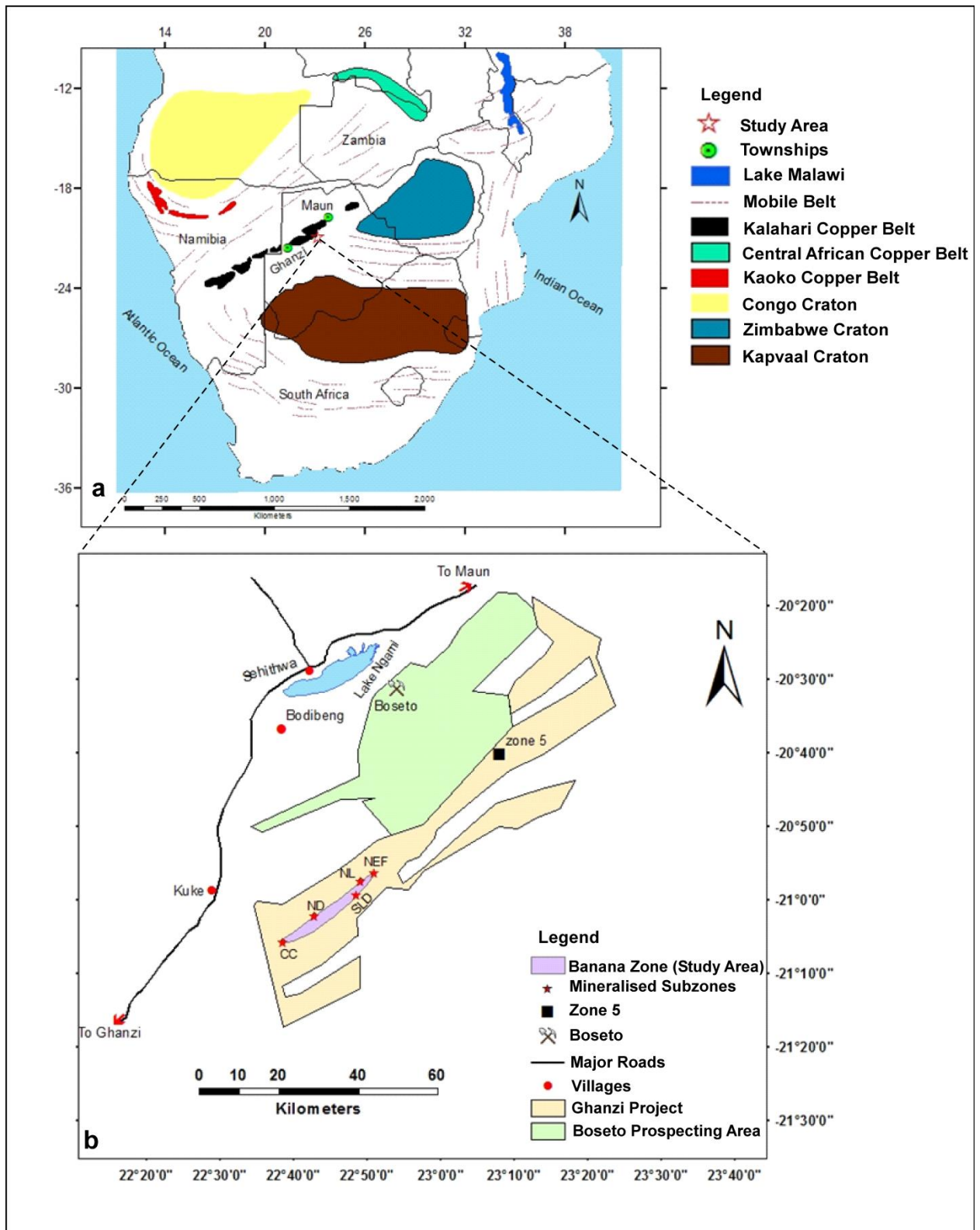


Fig. 1.1: (a). Schematic geological and structural map of southern Africa showing the location of the Ghanzi-Chobe Belt (GCB) and the Banana Zone in relation to the Congo, Zimbabwe, and Kaapvaal cratons (modified from **Maiden and Borg, 2011**). (b) Outline of the Banana Zone (red star in (a)) showing the different mineralised Subzones: CC=Chalcocite Zone, ND=New Discovery, NL=North Limb, NEF=North East Fold, SLD=South Limb Definition.

Table 1.1

Exploration history in the Ghanzi-Chobe Belt (GCB)

Year	Exploration Company	Activity
1962	Johannesburg Consolidated Investments (JCI)	Geological Mapping
1967	Anglo-Vaal – Joint Venture with De Beers, US Steel Corporation (US Steel) and Tsumeb Corporation	Drilling and soil geochemistry
1970- 1980	US Steel	Soil geochemistry, ground geophysics, trenching and drilling
1989- 1994	Anglo American Botswana	Airborne and ground geophysical surveys, Landsat and photo-geological interpretation, soil geochemistry and drilling
1996- 2000	Delta Gold (Delta) – Joint Venture with Kalahari Gold and Copper (Pty) Ltd (Kalahari Gold) and Gencor/Billiton	Deposit modelling
2005 to 2011	Discovery Metal Limited	Airborne and ground geophysical surveys, soil geochemistry and drilling. Mining activity 2011-2014.
2007 to present	Cupric Canyon Capital (formerly Hana Mining Limited) Acquisition of Boseto mine in 2015	Airborne and ground geophysical surveys, soil geochemistry and drilling
2011 to present	MOD Resources Limited	Airborne and ground geophysical surveys, soil geochemistry and drilling

1.2. Problem statement

All the previous geological studies and exploration works conducted in the KCB, namely the GCB in Botswana, mainly focused on the lithostratigraphy, depositional environment, and the mineralised horizon in a broad sense (see **Borg, 1988; Borg and Maiden, 1989; Modie, 1996a, 2000; Sillitoe et al., 2010; Hall, 2013**). A thorough description of the host rock petrology, geochemistry, alteration patterns, and the mineralised bodies is lacking and the provenance and tectonic setting of the sedimentary succession that hosts the Banana Zone Cu-Ag deposit is not well constrained. This is of significant importance considering that the chemical composition, the mineralogical composition, and the redox state may control metal precipitation and the patterns of hydrothermal alteration and therefore provide better information to understand the provenance and as well as the processes that converge to form the Cu-Ag mineralisation. Furthermore, ever since their discovery, the SHC deposits have been the second world's most important sources of copper (15% of world production; **Sillitoe, 2012**). In spite of these numerous exploration activities and enormous economic contributions of this area, the processes that combined to form the mineralisation are not yet well understood.

1.3. Aim and objectives of the study

The aim of this research is to constrain the petrogenesis of the Banana Zone Cu-Ag mineralisation and consequently understand the processes that combined to form the mineralisation in the GCB, particularly in the Banana Zone area. This will be achieved by:

- (i) Petrological and geochemical assessment of the host rocks in order to classify the rocks, evaluate the weathering intensity at the source area, and enhance the understanding of the sediments (siliciclastic) provenance and tectonic setting of the region.
- (ii) Characterisation of the style of mineralisation in the Banana Zone.
- (iii) Documenting the ore and hydrothermal alteration mineral paragenesis in the Banana Zone in order to arrive at the processes that could have led to the mineralisation.

- (iv) Proposing a comprehensive genetic model of the Banana Zone mineralisation that is likely to help in designing exploration models.

1.4. Methodology

The methods of study used to achieve the objectives of this research include field work at Khoemacau Copper Mining area and laboratory work carried out at the Council of Geoscience (COG), Pretoria South Africa and at the Botswana International University of Science and Technology (BIUST).

1.4.1. Fieldwork

Due to the extensive cover of Phanerozoic strata and scarcity of outcrops in the area, the investigations were conducted on drill cores made available by Khoemacau Copper Mining. Field investigations focussing on detailed lithology, structural features, alteration patterns, and mineralisation were carried out in approximately 4097 m of core from fifteen drill holes. Samples were collected from five locations within the Banana Zone, including: (i) the Chalcocite Zone (CC), (ii) New Discovery (ND), (iii) North Limb (NL), (iv) North East Fold (NEF), and; (v) South Limb Definition (SLD) (Fig. 1.1b; Appendix A1).

1.4.2. Laboratory work

Laboratory works involved numerous laboratory techniques including optical microscopy, X-ray diffraction (XRD), X-ray fluorescence (XRF), and inductively coupled plasma-mass spectrometry (ICP-MS).

A total of 176 samples, including 25 sandstone samples from the Ngwako Pan Formation and 30 sandstone samples, 83 siltstone and shale samples (considered together as mudrocks), and 38 carbonate rocks samples from the D'Kar Formation were selected for optical microscopy investigations. Standard petrographic polished thin sections were

studied using both transmitted and reflected light microscopes at the Botswana University of Sciences and Technology (BIUST). The modal composition data of twenty representative sandstone samples, including 10 sandstone samples from the Ngwako Pan Formation and 10 sandstone samples from the D'Kar Formation presented in Table 4.1 was determined visually using the standard charts of Terry and Chillingar (**Berkman, 2001**). Modal composition of the mudrocks and carbonate rocks of the D'Kar Formation could not be visually estimated/counted due to their fine grain size and/or significant carbonate content.

In order to compliment the petrographic observations, X-ray diffractometry (XRD) technique was employed. Sixteen samples were selected for XRD analyses. The analyses were performed at COG, Pretoria, South Africa using a Bruker D8 Advance, $\text{CuK}\alpha$, X-ray diffractometer (XRD) with the operating and measurement conditions following the method described by **Brime (1985)**. Megascopic data complimented by optical microscope studies and XRD data (Appendixes A2 and A3) were used to investigate and document the wall rock alteration, host rock mineralogy, and ore phases and reconstruction of the mineral paragenesis.

Ten of the least-weathered/altered samples of sandstone from Ngwako Pan and D'Kar Formations (five each), five samples each from the mudrocks and carbonates of the D'Kar Formation (Tables 4.1 and 4.2) were selected for geochemical investigations. The major element concentrations (expressed as oxides) were measured using XRF on fused beads at COG, Pretoria, South Africa. The fused beads were prepared by fusing 1 g of roasted sample and 10 g of flux consisting of 49.5 % $\text{Li}_2\text{B}_4\text{O}_7$, 49.5 % LiBO_2 , and 0.50 % LiI at 1150 °C. An in house amphibolite reference material (12/76) was used to check the data quality and the analytical accuracy. The analytical precision was determined based on the duplicated results and was generally better than 5 % for all elements.

Trace elements including REE abundances were determined by ICP-MS with a Perkin Elmer Elan 6000 at COG, Pretoria, South Africa. For trace element analyses excluding REE, 0.2 g of milled sample was digested in Savillex beakers with HCl, HNO_3 , HClO_4 and HF. The samples were dried and dissolved in HClO_4 . The drying process was repeated and then samples were re-dissolved in HNO_3 and H_2O_2 . For the REE analyses 0.5 g of sample was fused with 12:22 lithium tetraborate and lithium metaborate flux in Pt crucible and then dissolved in dilute HNO_3 . Standard samples GSD 09, GSD 10, GSD 11, GSR 1,

and GSR 3 were used for the calibration of trace elements. Accuracy of trace element analyses was within 5 %.

REE concentrations were normalised to Post Archean Average Australian Shale (PAAS, **Taylor and McLennan, 1985**) for the mudrocks and C1-chondrite (**Taylor and McLennan, 1985**) for the sandstones. For PAAS-normalised samples, REE fractionation was calculated as Pr_{SN}/Yb_{SN} (SN, shale normalised) avoiding anomalous La and Ce concentrations (**Franchi et al., 2016**). In order to avoid anomalous behaviour of La, Ce, Eu and giving the quasi-linear behaviour of the near neighbour REE in the logarithmic plots, REE anomalies were calculated using the geometric equation of **Lawrence et al. (2006)**. In chondrite normalisation, the Eu anomaly was calculated as $Eu/Eu^* = Eu_{CN}/((Sm_{CN}) \cdot (Gd_{CN}))^{1/2}$ (**Taylor and McLennan, 1985**). The subscript “CN” indicates Chondrite-normalised values. The Ce anomaly was defined as $Ce/Ce^* = Ce_{CN}/((La_{CN}) \cdot (Pr_{CN}))^{1/2}$ (**Kato and Nakamura, 2003**).

1.5. Thesis organisation

This thesis contains eight chapters. **Chapter 1** is the General Introduction. It covers the background of the study area, the location of the study area, as well as the statement of the problem. It also covers the aim and objectives of the study as well as the methodology used in this study.

Chapter 2 deals with the geological setting that includes both the regional geology and the local geology of the Banana Zone Cu-Ag mineralisation.

Chapter 3 documents the stratigraphic and petrographic characteristics of the sandstone of the Ngwako Pan Formation, and the carbonate rocks, the mudrocks, as well as the sandstone of the D’Kar Formation.

Chapter 4 deals with the geochemical characteristics of the sedimentary rocks spatially associated with the Banana Zone Cu-Ag mineralisation. The geochemistry component is divided into two sections. In the first section the rocks are characterised on the basis of their major elements content, whereas in the second section, rocks are characterised on the basis of trace elements.

Chapter 5 deals with the mineralisation style and mineral paragenesis of the Banana Zone.

Chapter 6 documents the hydrothermal alteration patterns and the associated ore mineral phases observed in these rocks.

Chapter 7 discusses the results obtained. This covers the classification of the sandstone of both the Ngwako Pan and D'Kar Formations, the provenance and source rocks characteristics, the weathering intensity of the source area, as well as the tectonic setting of the region. It also covers the paragenetic sequence, characteristics of the mineralising fluids, ore deposition stages, as well as the tentative genetic model.

Chapter 8 presents the general conclusions and the future work.

CHAPTER 2

GEOLOGICAL SETTINGS

This chapter deals with the geological setting that includes both the regional geology and the local geology of the Banana Zone Cu-Ag mineralisation.

2.1. Regional Geology

The Banana Zone study area is in northwestern Botswana, which is characterised by several Precambrian orogenic belts namely the Palaeoproterozoic Eburnian Magondi Belt, Mesoproterozoic Kibaran Ngamiland Belt, and Neoproterozoic to Early Palaeozoic Pan African Ghanzi-Chobe Belt (**Kampunzu et al., 2000**; Fig. 2.1). The Banana Zone Cu-Ag deposit is within the Ghanzi-Chobe Belt (GCB), which is part of the KCB in Botswana. The KCB stretches discontinuously from Namibia through Botswana, Zambia and into the Democratic Republic of Congo (**Pretorius and Park, 2011**). The KCB formed as a result of a collision between the Congo Craton and the Kalahari Craton during the amalgamation of Gondwana between 600 and 450 Ma (**Stanistreet et al., 1991**; Fig. 2.2).

The SW-NE trending GCB with a length of about 500 km and width of 100 km (**Borg and Maiden, 1989**; **Modie, 2000**; **Maiden and Borg, 2011**), extends from the Namibia-Botswana border in the south, to Shinamba Hills in the north, near the Botswana-Zambia border (**Kampunzu et al., 1998**; **Modie, 2000**), and has recently brought much attention to exploration companies, resulting in the successful identification of significant copper-silver resources (**Botswana Resource Sector Report, 2013**). Rocks of the GCB were deposited along the rifted northwest margin of the Kalahari Craton (**Master, 2010**). The GCB is characterised by a sequence of deformed meta-volcanics and meta-sediments, known in Namibia as the Sinclair Group (**Watters, 1977**), the Ghanzi-Chobe Supergroup in Botswana (**Modie et al., 1998**), and the Katanga Supergroup in Zambia (**Kampunzu et al., 1998**). These sedimentary rocks were deposited in an intracontinental rift followed by a proto oceanic rift system that finally evolved into a syn- to post-orogenic sequence (e.g., **Kampunzu et al., 1991**; **Miller et al., 2009**; **Master, 2010**).

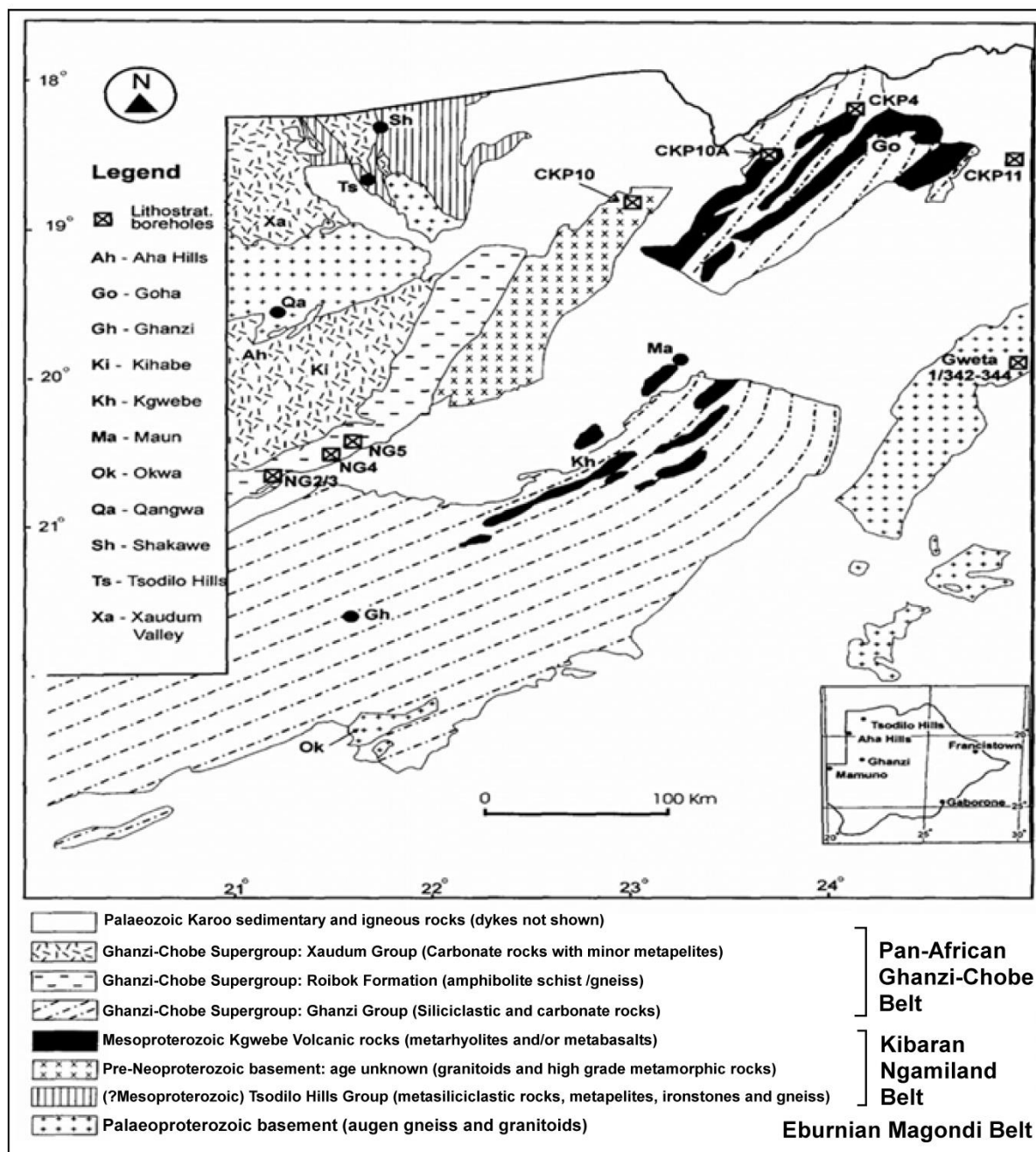


Fig. 2.1: Geological map of northern Botswana (From **Kampunzu et al., 2000**).

Northwestern Botswana is covered by an extensive cover of Phanerozoic strata, making the outcrops scarce. Therefore, investigation of the lithostratigraphy in the area remains very challenging. Nevertheless, aeromagnetic data, information from few available exposures, and borehole cores provides incredibly critical information useful in the lithostratigraphic study of the area. The summary of the geology of NW of Botswana as currently defined on the geological map of Botswana (**Kampunzu et al., 2000, and references therein**) and the main geological units are presented in Figure 2.1. Since the main lithostratigraphical units are commonly not exposed, the boundaries between units were inferred using aeromagnetic data (**Kampunzu et al., 2000**).

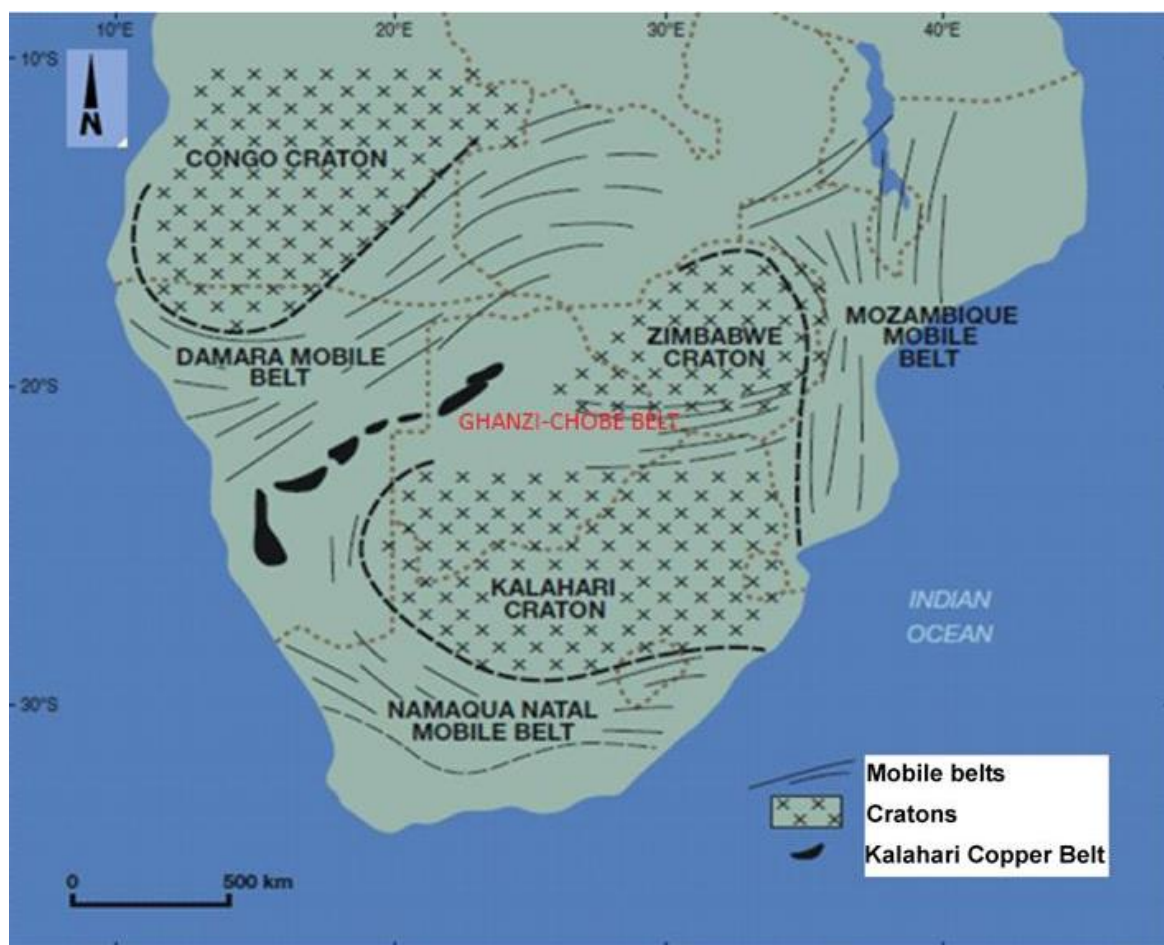


Fig. 2.2: The regional tectonic framework of southern Africa showing the position of the Ghanzi-Chobe Belt (From **Maiden and Borg, 2011**).

2.1.1. Palaeoproterozoic basement

Exposures of pre-Damara granitic and gneissic basement occur in the Okwa valley and Xangwa area (Fig. 2.1). The rocks exposed in the Okwa area are mostly porphyritic rhyolitic felsite, augen gneiss, granite, and microgranite (**Aldiss, 1988; Ramokate et al., 1996**). The Okwa gneiss has a strong foliation defined by biotite and feldspar with NE-SW trend and has a variable amount of mafic xenoliths of igneous origin. Basement rocks in the Okwa valley yielded U-Pb zircon crystallisation age of 2055.0 ± 1.3 Ma (**Ramokate et al., 1996; Mapeo et al., 2006**).

The exposures in the Xangwa area consist mainly of augen gneiss and granite that are faintly foliated to non-foliated (**Wright, 1958a**). The foliation in the augen gneiss is NE-SW trending with subvertical dip. U-Pb ages for zircons from the gneiss suggest a crystallisation age of 2.07 Ga (**Hanson et al., 1998**).

2.1.2. Mesoproterozoic basement

The Mesoproterozoic basement in NW Botswana is represented by the Kgwebe volcanic rocks. The Kgwebe volcanic rocks consist mostly of metabasalt and metarhyolite (**Modie, 1996b**). Several authors reported sandstone and conglomerate intercalations within the volcanic sequence, but the relationships between volcanic and sedimentary rocks are not clearly exposed and the sedimentary rocks might be slivers of Neoproterozoic sedimentary rocks tectonically juxtaposed to the metarhyolites (**Dietvorst and Gopolang, 1995**). Possible inliers of the Kgwebe volcanic rocks occur in the core of tight fold structures in the Goha and Shinamba Hills area (**Modie, 2000; Key and Ayres, 2000; Singletary et al., 2003**). U-Pb zircon dating of a sample of Kgwebe metarhyolite from Mabeleapodi Hills in the Ghanzi Ridge constrained the crystallisation age of the Kgwebe volcanic rocks at 1106 ± 2 Ma (**Schwartz et al., 1996; Modie, 2000; Kampunzu et al., 2000**).

The Kwando Complex is a gneiss-granitoid association, which is not exposed, and its inferred distribution in the northern and eastern part of the Okavango area is based on the interpretation of aeromagnetic data. Its composition has been documented in boreholes CKP-10 and CKP-10A (**Meixner and Peart, 1984**). The main rock type is pink granitic gneiss with abundant amphibole layers, which grades downwards into a hornblende-biotite–feldspar gneiss. The lithostructural evidence from the two boreholes shows that the

Kwando Complex includes syn-kinematic to late-kinematic batholiths (**Carney et al., 1994**).

2.1.3. Neoproterozoic metasedimentary units

The Neoproterozoic Ghanzi-Chobe Supergroup is formed by three lithostratigraphic units that include the Ghanzi Group, Roibok Complex, Kaonaka, Xaudum, and Tsodilo Hills Groups (**Kampunzu et al., 2000; Key and Ayers, 2000**). Moreover, the Ghanzi Group is unconformably overlain by the sedimentary rocks of the Okwa Group (**Kampunzu et al., 2000**). The relationships between the different Neoproterozoic units are still poorly constrained and these units are therefore independently described below.

2.1.3.1. Ghanzi Group

The best exposures of the Ghanzi Group are found in the vicinity of Mamuno and Kalkfontein settlements (**Ramokate et al., 2000**). The Ghanzi Group is a sequence of (meta-) sedimentary rocks that overlie the Kgwebe Formation with an unconformable contact (**Kampunzu et al., 1998**). The maximum thickness of the Ghanzi Group is about 13 km and is composed of red and green siliciclastic and carbonate rocks (**Ramokate et al., 2000 and references therein**) that are subdivided into four lithostratigraphic units namely; the Kuke Formation, the Ngwako Pan Formation, the D'Kar Formation, and the Mamuno Formation (**Modie et al., 1998**). These rocks have undergone lower greenschist facies metamorphism and tightly folded, with the regional SW-NE trend of fold axes.

In the Shinamba Hills copper-bearing sedimentary rocks occur and are tentatively equated with the mineralised rock successions of the Ghanzi Group. They consist of massive to bedded limestones, calcareous siltstones, and copper-bearing grey siliceous limestones. They have been tightly folded and contain a prominent axial planar cleavage trending ENE-WSW (**Wright, 1958a; Thomas, 1969**).

2.1.3.2. Roibok Complex

The Roibok Complex is composed of schist and gneissic rocks that are not exposed. These rocks were encountered in cored boreholes located in the Roibok Laagte dry valley. (**Ludtke et al., 1986; Singletary et al., 2003**) described the cores as a succession of amphibolites intercalated in leucocratic magnetite-bearing schists and gneisses with

numerous quartzose layers. They form a prominent magnetic marker believed by **Ludtke et al. (1986)** to be an extension of the Matchless Amphibolite Member of the Damara sequence in Namibia. The core samples are invariably highly foliated and contain zones of enhanced shear deformation in which the fabric intensifies and porphyroblasts become rotated. Foliation attitude determined by down-hole probes shows dip of 45-80° SE.

2.1.3.3. Kaonaka Group

Dolomitic marbles are exposed in the Kihabe and the Kaonaka Hills. These dolomitic marbles can either be massive, coarsely foliated or recrystallised and they contain dolomite and subsidiary ankerite, siderite, and sericite or talc (**Wright, 1958b**). Tremolite, wollastonite, and garnet were also reported by **Vermaak (1962)**. He indicated that the dolomitic marbles have been folded into isoclinal structures with axial plane dipping to the southeast. The dolomitic marble contains chalcopyrite in the southeast of the Kaonaka Hills (**Wright, 1958c**). The most important structural features consist of south to southwest plunging rodding lineation defined by quartz reefs, ankerite lenses, and flakes of talc or sericite.

2.1.3.4. Xaudum Group

The Xaudum Group is composed of a varied sequence of metasedimentary rocks. The rocks occur into two settings, which possibly represent separate structural domains. The southeastern succession is mainly composed of dolomite and limestone with intercalations of chert exposed in the Aha Hills. The second sequence occurs in the Xaudum Valley to the north. It contains a greater variety of sedimentary rock types including: shale, sandstone, dolomite and limestone similar to those in the Aha Hills. The cherty carbonate occurrences in the Xaudum Valley overlie the mixed assemblage of quartzitic sandstone and shale. Sideritic limestone observed to the east of Xaudum Village is brown coloured and “microconglomeratic” and appears to cap the cherty carbonate unit (**Lemaire, 1971**). The Xaudum Group rocks are younger than 1020 Ma, based on the age of detrital zircon from sandstone near Shakawe (**Mapeo et al., 2000**).

2.1.3.5. Tsodilo Hills Group

The Tsodilo Hills Group is characterised by an association of ferruginous and micaceous metamorphosed rocks. These rocks that are exposed in Tsodilo Hills include quartzite, quartz-mica schist, metamorphosed conglomerate, minor shale, phyllite, sandstone, and ironstone (**Singletary et al., 2003; Wendorff, 2005**). Furthermore, these rocks which extend to the Shakawe area have been altered to kyanite metamorphic grade (**Wendorff, 2005 and references therein**). The metamorphic muscovite from the Tsodilo Hills yield $^{40}\text{Ar}/^{39}\text{Ar}$ age of 490.0 ± 2.3 Ma. This age, according to **Singletary et al (2003)**, implies that these rocks were involved in the Pan-African orogenic movements.

According to **Vermaak (1962)**, the Tsodilo Hills Group includes four informal lithostratigraphic units including from bottom to top:

- 1) lowermost unit of coarse- to fine-grained quartzite, itabiritic quartzite and micaceous quartzite overlain by;
- 2) micaceous and itabiritic quartzites intercalated with mica schist;
- 3) current-bedded quartzites and micaceous quartzites with conglomerate and;
- 4) an uppermost unit made of itabiritic quartzite with intercalated pebble and micaceous bands.

2.1.4. Karoo Supergroup

The Karoo Supergroup comprises a succession of sedimentary and volcanic rocks that unconformably overlie Archaean and Proterozoic beds and are overlain unconformably by the Kalahari beds (**Carney et al., 1994**).

At the base of the Karoo Supergroup, the Dwyka Group comprises diverse lithofacies deposited during a major Permo-Carboniferous glaciation. The Dwyka Group is overlain by the Eccra and Beaufort Groups, made up of shallow marine to fluvial sedimentary rocks that reflect conditions of increasing continentality and a change to warmer, more arid palaeoclimates. The Karoo sedimentation ended in the Lower Jurassic with outpourings of tholeiitic flood basalts of the Stormberg Lava Group onto a desert landscape formed of aeolian sands of the Lebung Group (**Carney et al., 1994**).

2.1.5. Kalahari Group

The Stormberg Basalts are unconformably overlain by a thin veneer of loosely consolidated Cainozoic to Recent deposits termed the Kalahari Group. The deposits are widespread and cover almost 90 % of the GCB (**Modie, 1996a**). The Kalahari beds consist of sediments of a wide variety, collected in aeolian, fluvial, and lacustrine environments. Although there is no definite stratigraphic subdivision of this unit, **Jones (1980)** gave a lithological summary of the Kalahari beds. These lithologies include marls or clays with pink or red colour, poorly sorted gravels; aeolian sands characterised by reddish-brown or buff to whitish colour and are fine-grained and finally duricrusts comprising calcretes, silcretes, and ferricretes.

2.2. Local Geology

2.2.1. Lithostratigraphy

Rocks of the Ghanzi-Chobe Fold Belt are exposed along the northeast-trending Ghanzi Ridge, which extends from the Botswana-Namibian border towards Lake Ngami, and areas of the Goha and Shinamba Hills (**Pretorius and Park, 2011; Lehmann et al., 2015**). The base of the Ghanzi Ridge sequence is composed of the Kgwebe Formation, which is unconformably overlain by a younger Ghanzi Group (**Thomas, 1969, 1973; Modie et al., 1998; Modie, 2000**). **Modie et al. (1998)** formally subdivided the Ghanzi Group into four formations namely the Kuke, Ngwako, D'Kar, and Mamuno Formations in ascending stratigraphic order (Table 2.1).

Table 2.1

Stratigraphy of the Ghanzi-Chobe Supergroup (modified after Modie et al., 1998).

Age	Group	Formation (thickness)	Lithology	Depositional setting
NEOPROTEROZOIC	GHANZI	MAMUNO (~ 1500 m)	Arkose, siltstone, shale, mudstone and limestone	Alluvial system changing to shallow marine and progradational shoreline
		D’KAR (~ 1500 m) Cu-Ag mineralization	Shale, mudstone, siltstone, arkose, marlstone and limestone	
		NGWAKO PAN (~ 2000 m)	Arkose, sub-arkose, siltstone and shale	
		KUKE (~ 500 m)	Sandstone and conglomerate	
MESOPROTEROZOIC	KGWEBE VOLCANICS (~ 2000 m)		Metarhyolite, tuff, metabasalt, agglomerate, minor intercalated metasedimentary rocks	Bi-modal volcanism

2.2.1.1. Kgwebe Formation

Exposures of the Kgwebe Formation volcanic rocks have been found in the core of a regional anticline in the Mabeleapodi and Kgwebe Hills. They trend north-east of the village of Kuke (**Pretorius and Park, 2011**). They are the oldest exposed structural and stratigraphic units within the Project area. This sequence with a composite thickness of approximately 2000 m is dominated by bimodal volcanic rocks of porphyritic rhyolite (Fig. 2.3) and dacites, pyroclastic flow deposits, minor peperites, and sub-aerial basaltic lavas with subordinate flow breccias, which are intercalated with minor epiclastic and

tuffaceous sedimentary rocks and sandstones (**Kampunzu et al., 1998; Modie, 2000**). According to **Kampunzu et al. (1998)** these volcanic rocks represent a bimodal suite of within-plate low titanium-phosphorus continental tholeiites and post-orogenic high-K rhyolites emplaced during a collision-related extensional collapse event. **Modie (1996a)** proposed that this formation has been deposited in the initial rifting stage of an intra-continental rift basin. Post-volcanism thermal subsidence and down-sagging of the basin along bounding faults resulted in the deposition of fluvial and volcanoclastic sediments above the volcanic rocks. **Schwartz et al. (1996)** obtained a zircon U-Pb age of 1106 ± 2 Ma from the basal rhyolite at Mabeleapodi Hills. Additionally, this age according to **Hegenberger and Burger (1986)** is within the error of the age of 1094 ± 20 Ma for the Oorlogsende Quartz Porphyry situated some 60 km from the Botswana border in eastern Namibia. The age of the Kgwebe Porphyry is the only reliable depositional age for any lithological unit within the GCB. Therefore, it represents the maximum age of the Ghanzi Group sedimentary deposits.



Fig. 2.3: Porphyritic rhyolite of the Kgwebe Formation, showing characteristic medium-grained feldspar phenocrysts (white) set in a fine-grained matrix of interlocking crystalline texture (from **Pretorius and Park, 2011**).

2.2.1.2. Ghanzi Group

The relationships between tectonic and primary structures (e.g., folds, cleavage, igneous foliation, bedding and cross-bedding in sedimentary rocks) show that the Ghanzi Group overlies unconformably the Kgwebe volcanic rocks (**Kampunzu et al., 1998**). This group, which comprises a continental sedimentary succession, have been deposited during a marine transgression. The deposition of the Ghanzi Group is bracketed between 1047 ± 24 Ma (youngest detrital zircons; **Kampunzu et al., 2000**) and 530 Ma (the date for peak metamorphism; **Ahrendt et al., 1978**). Although terms applied to sedimentary rocks are used to describe these rocks of the GCB, the formations are distinctly metamorphic.

The Kuke Formation with a thickness of a 500 m forms the base of the Ghanzi Group and rests unconformably above the Kgwebe Formation (**Modie, 2000**). It consists of grey quartz arenites and thin red sandstone with basal conglomerate and pebbly sandstone. Included in the conglomerate are clasts of the underlying Kgwebe Volcanics, suggesting an input from extra-basinal source (**Modie, 2000; Kampunzu et al., 2000**). The Kuke Formation is characterised by cross-bedding, graded bedding, reactivation surfaces and pebbly layers (**Modie, 2000**). Hitherto this unit has been assigned to the Kgwebe Formation.

Overlying the Kuke Formation is the Ngwako Pan Formation. The best exposures of the Ngwako Pan Formation are found in the Ghanzi region. This formation that represents a succession deposited in an oxic environment is approximately 2,000 m thick in the region of the Ghanzi Ridge to 3,500 m below cover to the NE (**Modie, 2000**). The Ngwako Pan Formation includes poorly sorted, matrix-rich grey sandstones with purplish-red mudstone intercalations at the base and moderate to well- sorted reddish arkosic sandstone, minor siltstone and red mudstone at the top of the succession. The Ngwako Pan Formation is characterised by graded bedding, bipolar cross-bedding and pebbly layers. The uppermost 10-40 meters of the Ngwako Pan Formation contains abundant pebble-rich beds throughout the Ghanzi Ridge area. These beds have been interpreted as high-energy fluvial deposits formed in response to renewed basin subsidence (**Modie, 1996b**). **Master (2010)** described hummocky cross-stratification in association with very thin drapes of shale. Furthermore, he observed that surfaces with low angle truncations and gentle curves resembling swaley cross-stratification are common in plane-laminated sandstones. **Modie (1996b)** regarded the rocks of the Ngwako Pan Formation to have been deposited in an

axial fluvial system, although **Master (2010)** and **Caterall et al. (2012)** suggested deposition occurred within lower, middle, and upper shoreface environments adjacent to alluvial and fluvial sources.

The Ngwako Pan Formation is conformably overlain by a ca. 1500 m thick marine siliciclastic and carbonate package of D’Kar Formation. This formation that indicates a succession deposited in a reduced evaporitic environment marks a significant transgressive event in the Ghanzi-Chobe Basin (**Modie, 1996b**). In addition, this transgression induced the deposition of very fine-grained siliciclastic rocks with carbonates at the base and fine-grained sandstones at the top. The siliciclastic rocks are characterised by a repeated fining-upward succession of pale siltstones interbedded with dark shale beds suggesting a rhythmic seasonal sedimentation due to storm activities (**Modie, 1996b**). These rocks have a high content of fine-grained pyritic organic material and occasionally high carbon content leading them to be classified as black shales. Moreover, the occurrence of carbonate rocks in the lower D’Kar Formation throughout the area suggests deposition in shallow, warm waters of restricted lagoons or playa lakes (**Modie, 1996b**). Planar parallel lamination and cross lamination, ripple marks and algal rip up clasts are the dominant structures characterising these lithologies.

The upper succession of the Ghanzi Group is occupied by the Mamuno Formation. The Mamuno Formation conformably overlies the upper D’Kar Formation with gradational contact. The Mamuno is about 1,500 m thick and mainly composed of fine-to medium-grained reddish sandstone, siltstone and mudstone with subordinate limestone. The sedimentary structures are dominated by cross-bedding and wave ripples. The association of the red bed facies with alluvial-fluvial systems marks the transition between continental systems and the marine environment (**Modie, 1996a**).

2.2.2. Tectonic evolution

The tectonic evolution of the region commenced with the formation of rift basins along the northwest margin of the Kalahari Craton, initiated by extensional tectonic environment related to the Rodinia break up (**Modie, 1996a, b**). Additionally, this extensional tectonics led to intracontinental rifting and subsequent bimodal volcanism about 1106 Ma ago. During this time, sedimentation jointly with volcanism resulted in deposition of volcanoclastic of the Kgwebe Formation. The presence of primary structures (e.g., cross-

bedding, small-scale ripple marks, mudstone intraclasts, desiccation cracks and finning upward lamination pluses) indicate fluctuations between sedimentation and reworking at shallow water to emergent conditions (**Modie, 1996a, 2000**).

Renewed rifting and subsidence after volcanism led to the development of major sedimentary basins, i.e., the Passarge and Nosop basins (**Singletary et al., 2003 and references therein**). These basins accommodated the Ghanzi Group commencing with the deposition of the Kuke Formation near the rift margins.

Crustal rapturing *ca.* 770 Ma resulted in the emplacement of tholeiitic basalts and mafic dykes with composition of mid-oceanic rift basalts (MORB) belonging to the Roibok Formation. This formation can be traced into Namibia and has been correlated with the Matchless Amphibolite (**Ludtke et al., 1986**).

A long term, slow regional subsidence led to further enlargement and extension resulting in the establishment of shelf and basin environments, accompanied by marine transgression. In these environments accumulated about 5000 m thick siliciclastic and minor carbonates rocks of the Ngwako Pan, D'Kar, and Mamuno Formations (**Modie, 1996b; 2000**). Sedimentation in the basin was cyclical, alternating between carbonate- and clastic- dominated layers.

Although no age determinations are available for these units, **SACS (1980)** believed that deposition would have taken place between 950 and 1000 Ma. Rifting was aborted before the development of an ocean and hence resulted in a failed intracratonic rift system. During the assembly of the Gondwana Supercontinent (~550 Ma) changes in stress regime from extensional to compressional, led to the collision between the Congo and Kalahari Cratons and the formation of the GCB (see **Van Langendonck et al., 2013**). Subsequent deformation during the Pan African Damaran Orogeny resulted in regional-scale folding, faulting, and metamorphism of these rocks to lower greenschist facies (**Schwartz et al., 1995**). The pelitic rocks developed a slaty cleavage, and the arenites were deformed by brittle fracturing.

CHAPTER 3

STRATIGRAPHY AND PETROGRAPHY OF THE BANANA ZONE

Out of the four lithostratigraphic units constituting the Ghanzi Group (see Chapter 2), only two units have been intersected by drilling in the Banana Zone. These units include the oxidised, but locally reduced footwall sandstone of the Ngwako Pan Formation and the reduced shallow marine to lacustrine facies of the D’Kar Formation. Though the rocks have undergone lower greenschist facies metamorphism during the Damaran-Pan African orogenic event, delicate sedimentary structures, such as algal mats, are still preserved, hence the use of the names of the lithotypes without the prefix “meta”.

3.1. Ngwako Pan Formation

The Ngwako Pan Formation, the stratigraphically lowest unit intersected by drilling in the Banana Zone, is arenaceous and argillaceous and composed of sandstone interbedded with mudrocks (siltstone and mudstone). This unit is typified by its reddish brown to a pinkish colour, and occasionally greyish due to its high content of carbonate minerals. In most cases rocks from the Ngwako Pan Formation are massively bedded (Fig. 3.1a), nonetheless locally planar laminations marked by alternating argillaceous and arenaceous layers (Fig. 3.1b) and heavy minerals (e.g., rutile, zircon, magnetite, and ilmenite) (Fig. 3.1c,) are observed. The Ngwako Pan Formation is also characterised by graded bedding (Fig. 3.1d).

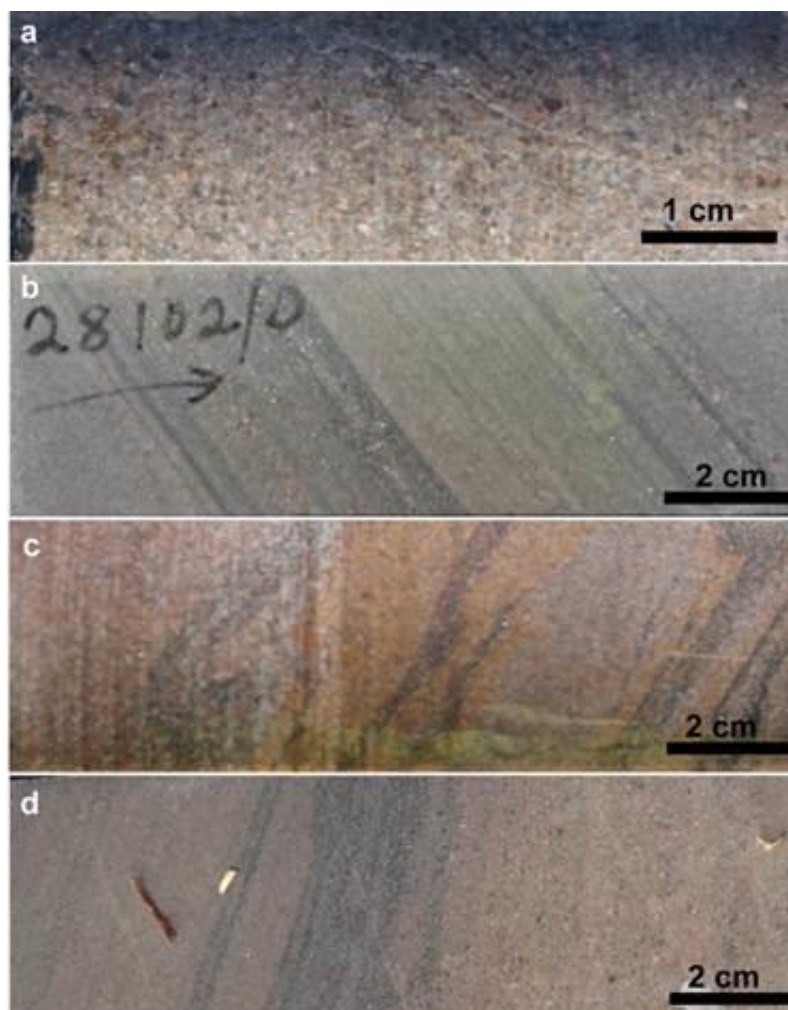


Fig. 3.1: Photographs of sandstone of the Ngwako Pan Formation illustrating variations in colour, texture and structures. (a) Very coarse grained massively bedded pink sandstone (drill core HA 17-D; 256.73-256.80 m). (b) Medium grained, planar laminated greyish brown sandstone (drill core HA 526-D; 165.00-165.30 m). (c) Very fine grained, finely laminated yellowish brown sandstone (drill core HA 17-D; 35.88-35.99 m). (d) Very fine grained, faintly laminated brown sandstone characterised by graded bedding (drill core HA 38-D; 159.36-159.42 m).

Petrographic study of the sandstone of the Ngwako Pan Formation reveals that the sandstone is poorly- to moderately sorted, consisting of fine sand size clasts to very coarse (pebbly) sandstone and micro-rudite (0.125 to 2.30 mm in size). Finer grains are generally angular to sub angular, whereas the medium to coarse-grained clasts are subrounded to well-rounded. This sandstone is texturally immature to submature, as indicated by the poor sorting and the angularity of the grains. The studied samples are mainly composed of quartz (Qt), feldspar (F), and lithic fragments (L). Trace amounts of detrital biotite and Fe-

(Ti) oxides are also observed. The average compositional ratio is $Q_{t64}: F_{32}: L_4$, and falls within the arkose field (Fig. 3.2), suggesting its compositional immaturity.

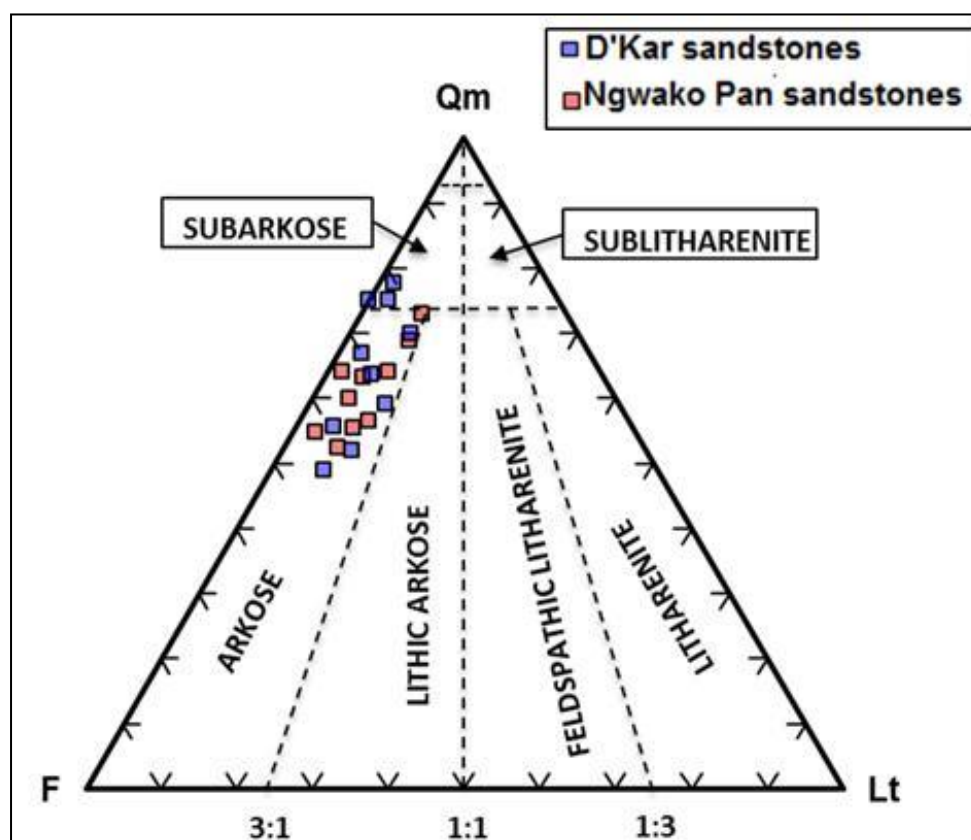


Fig. 3.2: Ternary plot for classification of sandstones (after **Folk, 1980**). Qm = monocrystalline quartz, F = Total feldspar grains (Plagioclase + K-feldspar), Lt = Total siliciclastic lithic fragments.

Quartz is the predominant framework grain and occurs as both mono- and polycrystalline quartz. Monocrystalline quartz is the most common type ranging from 52.5 to 73.0 vol. %, whereas polycrystalline quartz is present in a relatively small amount ranging from 0.0 to 6.1 vol. % (average 2.5 vol. %) (Table 3.1). Most of the monocrystalline quartz grains are inclusions-free and characterised by non-undulose extinction (Fig. 3.3a). Locally, quartz grains in the medium to coarse sand size range exhibit undulose extinction. Four types of polycrystalline quartz grains were observed: (i) grains composed of few equant crystals with no or little intercrystalline suturing (Fig. 3.3a); (ii) grains composed of crystals with both sharp and sutured boundaries (Fig. 3.3b); (iii) grains composed of polygonal crystals intersecting at an angle of 120° (Figs. 3.3c) and (iv) recrystallised metamorphic quartz grains (Figs. 3.3d).

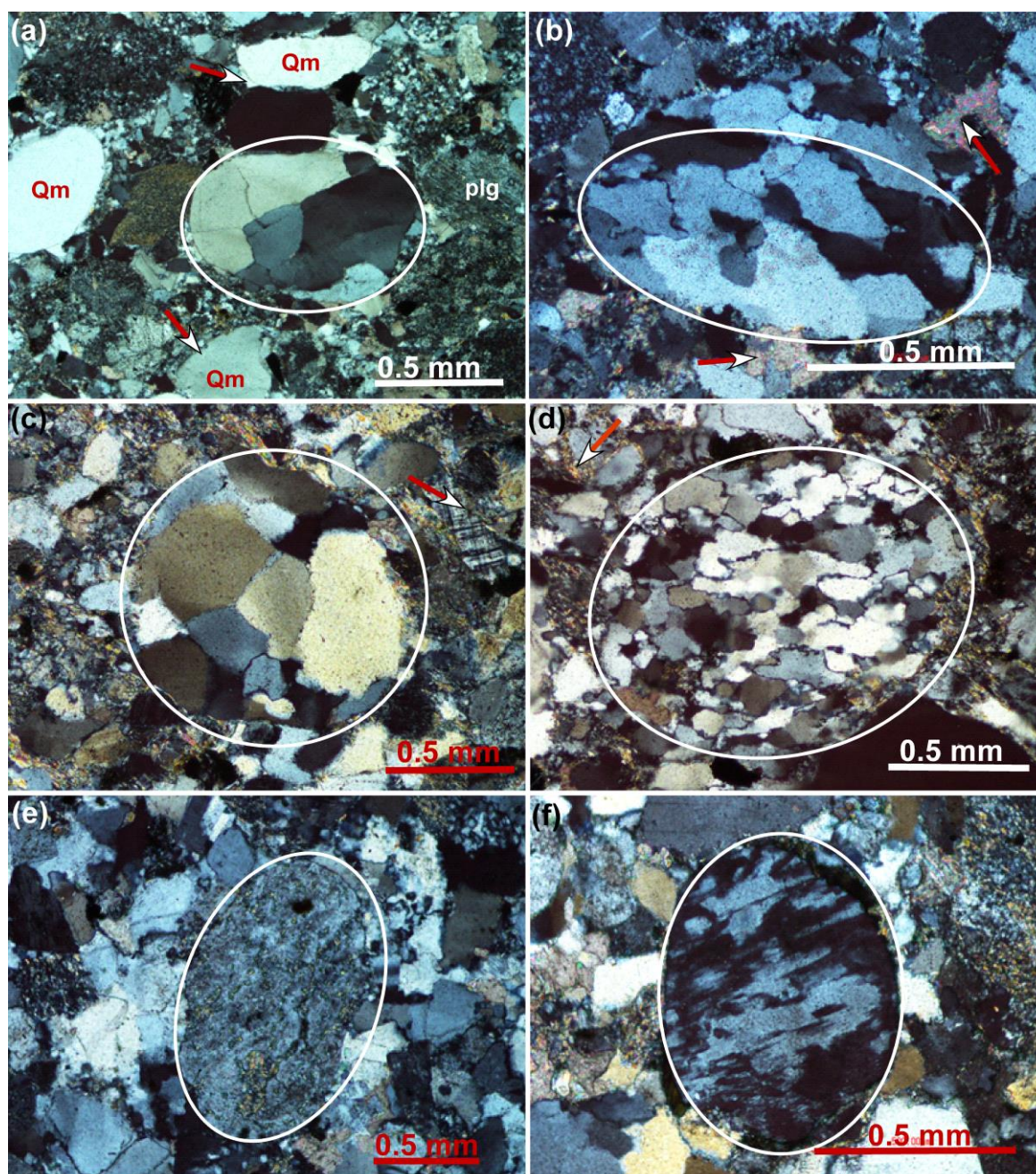


Fig. 3.3: Photomicrographs (transmitted light) of different types of polycrystalline quartz grains (white circles), monocrystalline quartz grains, feldspar grains and different cements. All photomicrographs are in cross polarized light. (a) Polycrystalline quartz of probable igneous origin; the grain is composed of few equant crystals with little or no intercrystalline suturing; quartz cement (arrow); monocrystalline quartz (Qm) grains with non-undulose extinction and partially altered plagioclase (plg) grain. (b) Polycrystalline quartz of probable metamorphic origin; the grain consists of crystals with both sharp and sutured boundaries; calcite cement (arrows). (c) Polycrystalline quartz grains of probable metamorphic origin with polygonal crystal boundaries at 120° ; Microcline (arrow) (d) Polycrystalline quartz grain of probable metamorphic origin with clear evidences of recrystallisation; Clay minerals (arrow). (e) Partially altered orthoclase grain; quartz cement (arrow). (f) Perthite grain.

The studied sandstone is rich in feldspar (19.0-42.1 vol. %; average 32.4 vol. %, Table 3.1), which is either unaltered or heavily weathered and altered to argillite, sericite or carbonate. K-feldspar (average 19.2 vol. %) is the most abundant feldspar and it includes microcline (Fig. 3.3c), orthoclase, and microperthite (Fig. 3.3e, f). Muscovite, heavy minerals (magnetite, zircon, and rutile), and myrmekite occur in minor quantities, but are invariably conspicuous and important framework components.

Lithoclasts are dominated by mudstone fragments. Other lithoclasts found include recrystallised chert, quartzite, schist, granitic, and volcanic lithic fragments (Fig. 3.4).

The amount of matrix is highly variable and is always below 15 vol. % modal. The matrix is composed of phyllosilicates (chlorite, sericite, and muscovite) and silt size clasts of quartz and feldspar. In addition to the matrix, the grains are cemented by mostly sparry calcite (that is fine-grained and without twinning) and locally by quartz that grows in crystallographic and optical continuity with the quartz clasts. Early iron oxide grain-coating cement is commonly preserved in well-developed quartz overgrowths.

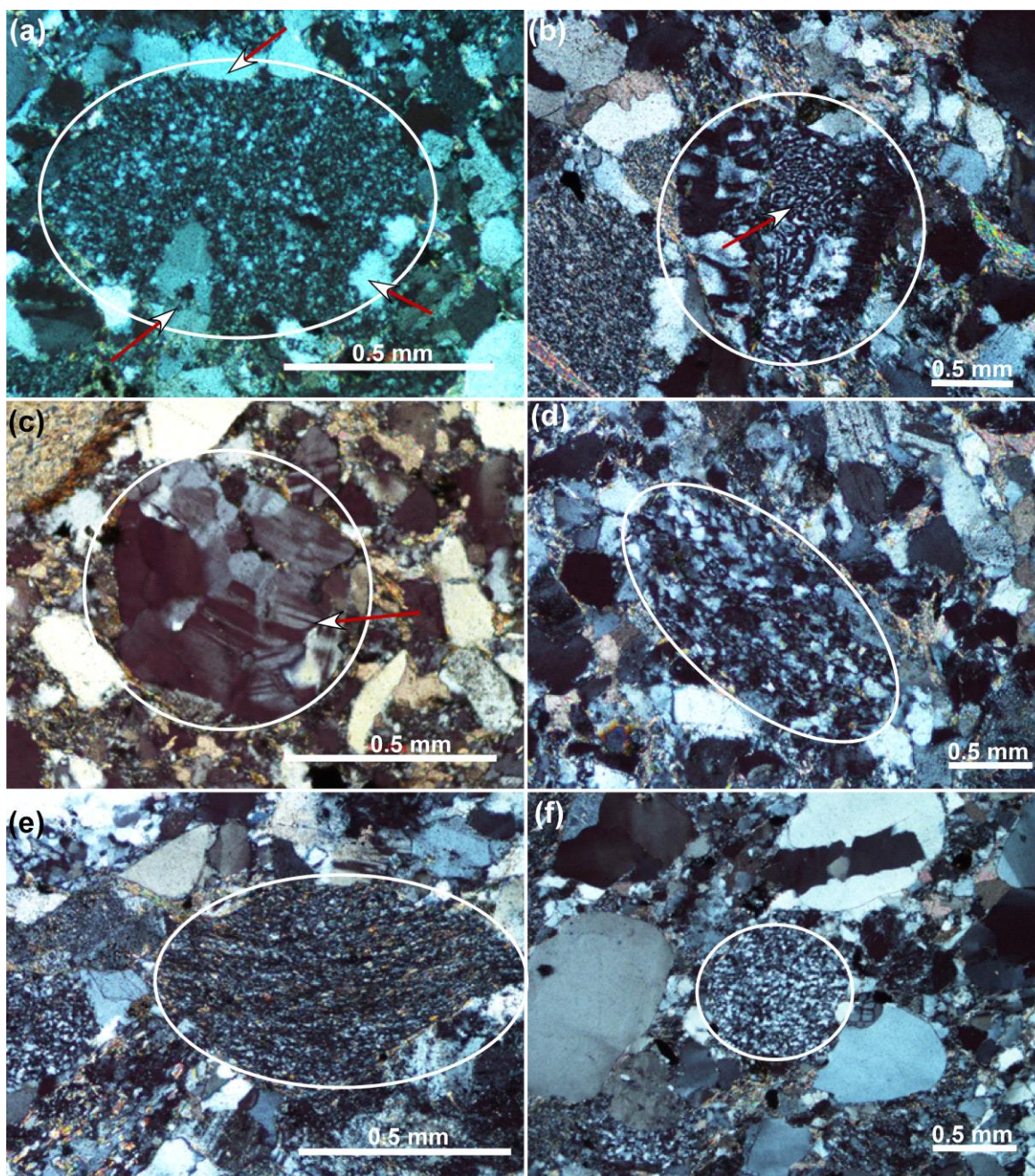


Fig. 3.4: Photomicrographs of lithic grains (white circles). All photomicrographs are in cross polarized light. (a) Volcanic rock fragments containing K-feldspar grains (arrows). (b, c) Fragments of a granitoid rock; myrmekitic intergrowth (arrow) can be recognized in (b) and twin lamellae (arrow) in (c). (d) Fragment of quartzite. (e) Mudrock fragment (f) Recrystallised chert fragment.

Table 3.1

Detrital modal compositions of selected sandstone samples from the Ngwako Pan and the D'Kar Formations

Sample No.	Qt	F	L	Qm	F	Lt	Qp	Lm	Li	Ls	Qm	P	K
CS1	68.9	22.7	8.4	68.9	22.7	8.4	20.0	0.0	0.0	80.0	73.8	9.2	17.0
CS2	57.9	36.8	5.3	55.4	36.8	7.8	66.7	0.0	0.0	33.3	58.7	16.3	25.0
CS3	58.6	40.4	2.0	52.5	40.4	7.1	77.8	0.0	0.0	22.2	55.6	16.7	27.7
CS4	65.7	34.3	1.0	64.6	34.2	2.0	50.0	0.0	0.0	50.0	65.9	10.1	24.0
CS5	76.0	19.0	5.0	73.0	19.0	8.0	78.8	9.1	6.1	6.1	75.0	11.9	13.1
CS6	65.0	35.0	0.0	60.0	35.0	5.0	100.0	0.0	0.0	0.0	63.2	15.7	21.1
CS7	66.3	31.6	2.1	63.2	31.6	5.2	60.0	20.0	0.0	20.0	66.7	14.4	18.9
CS8	58.6	34.3	7.1	56.5	34.3	9.2	45.0	10.0	10.0	10.0	60.5	14.2	25.3
CS9	64.0	28.0	8.0	64.0	28.0	8.0	52.4	9.5	0.0	38.1	67.8	11.4	20.8
CS10	56.8	42.1	1.1	54.7	42.1	3.2	66.7	0.0	0.0	33.3	56.5	18.5	25.0
Average	63.8	32.4	4.0	61.3	32.4	6.4	61.7	4.9	1.6	29.3	64.4	13.8	21.8
MS1	79.6	20.4	0.0	77.6	20.4	2.0	100.0	0.0	0.0	0.0	79.2	8.4	12.4
MS2	58.6	39.4	2.0	55.6	39.4	5.0	60.0	0.0	0.0	40.0	58.5	26.6	14.9
MS3	50.0	43.9	6.1	48.9	43.9	7.2	30.0	15.0	5.0	50.0	51.1	17.0	31.9
MS4	54.0	38.8	7.2	51.9	38.8	9.3	40.0	50.0	10.0	0.0	55.8	15.7	28.5
MS5	75.0	22.0	3.0	70.0	22.0	8.0	62.5	11.5	1.0	25.0	76.1	6.5	17.4
MS6	65.7	30.3	4.0	63.6	30.3	6.1	33.3	12.7	4.0	50.0	67.7	14.0	18.3
MS7	64.3	30.6	5.1	59.2	30.6	10.2	50.0	20.0	0.0	30.0	65.9	13.6	20.5
MS8	75.0	25.0	0.0	75.0	25.0	0.0	0.0	0.0	0.0	0.0	75.0	15.0	10.0
MS9	70.0	30.0	0.0	67.0	30.0	3.0	100.0	0.0	0.0	0.0	69.1	12.3	18.6
MS10	75.0	22.5	2.5	75.0	22.5	2.5	0.0	0.0	0.0	100.0	76.9	7.7	15.4
Average	66.7	30.3	3.0	64.4	30.3	5.3	47.6	10.9	2.0	29.5	67.5	13.7	18.8

CS = Ngwako Pan Formation sandstones; MS = D'Kar Formation sandstones; Qt = Total quartz (Qm + Qp); Qm = Monocrystalline quartz; Qp = Polycrystalline quartz; F = Total feldspar grains (P + K); P = Plagioclase grains; K = K-feldspar grains; L = Unstable lithic fragments (Lm + Li + Ls); Lm = Metamorphic lithic fragments; Li = Igneous fragments; Ls = Sedimentary fragments; Lt = Total lithic fragments (L + Qp). After **Dickinson (1985)**.

3.2. D’Kar Formation

The D’Kar Formation conformably overlies the Ngwako Pan Formation usually with a sharp and occasionally gradational contact. This unit which marks a transition to a marine or lacustrine environment (Modie, 1996) is composed of carbonate rocks, mudrocks and sandstone.

Carbonate rocks of the D’Kar Formation are composed of pinkish-white crystalline limestone (Fig. 3.5a) and yellowish grey or greyish black marlstones characterised by lamination that is marked by undulating marly and silty layers that give the rock a “zebra-skin” appearance (Fig. 3.5b). Occasionally, the marlstone is strongly sheared (Fig. 3.5c) and folded. Petrographic study of the limestone shows calcite crystals that are ribbon-like and aligned with their long axis parallel to the fabric. Similarly, petrographic study of the marlstones shows that the marly rich layers (0.5-0.7 mm thick) are composed of elliptically shaped calcite crystals that are aligned parallel to the laminations, and detrital silicate grains while the silty layers (0.2-0.6 mm thick) are composed of quartz, feldspar, muscovite, and chlorite.

Mudrocks of the D’Kar Formation are composed of green to grey siltstone and mudstone and/or shale. The siltstones are very fine-grained and finely laminated and locally disturbed by syn-sedimentary faulting (Fig. 3.6a). They consist essentially of silt-size (0.02-0.06 mm) angular quartz and subordinate feldspar, aligned muscovite, sericite, and chlorite flakes in a calcareous matrix (Fig. 3.7a). Monocrystalline quartz with undulose extinction is the most common quartz phase. The mudstone and/or shale are finely laminated and consist of elliptical quartz in a muscovite and chlorite-rich matrix (Fig. 3.7b) and exhibits microbial mats (Fig. 3.6b). The quartz and chlorite grains have their long axes aligned parallel to the lamination. Locally, the shale contains high amounts of organic matter, giving the rock its black colour. Also, many nodules, lenses and bands of dolomite are observed within the shale (Fig. 3.6c). Isolated patches of evaporite occurrences are observed within the siltstone and shale units (Fig. 3.7c, d)

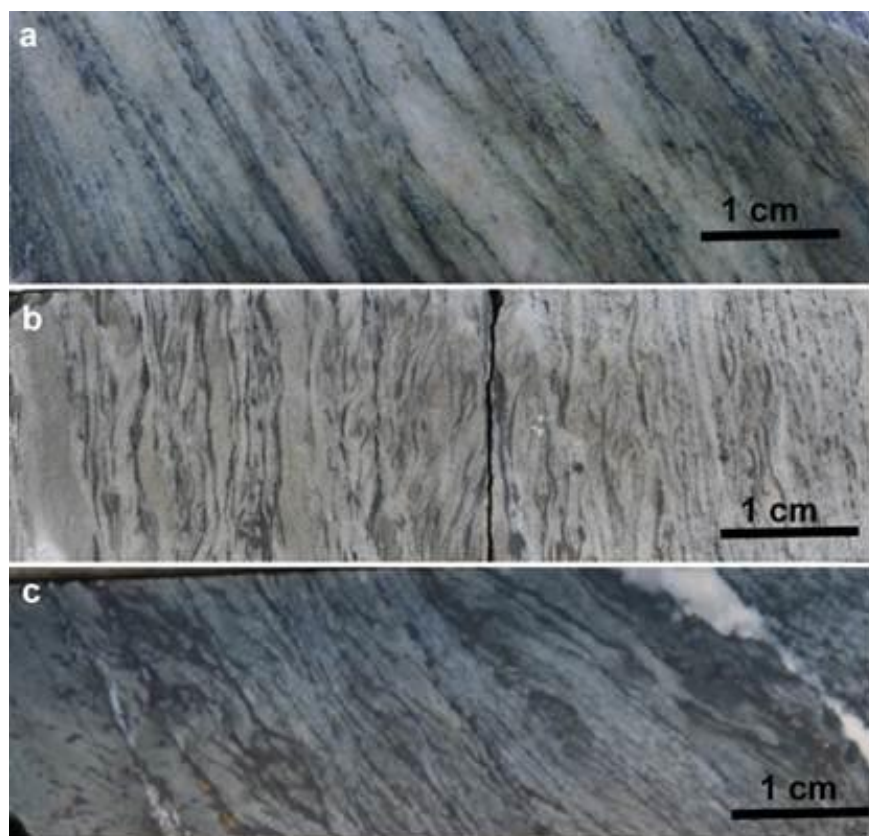


Fig. 3.5: Photographs of carbonate rocks of the D'Kar Formation. (a) Limestone hosting disseminated bornite (drill core HA 515-D; 202.78-202.92 m). (b) Marlstone characterised by undulating black and white strips ‘zebra skin appearance’ (drill hole HA 10-D; 169.27-169.40 m). (c) Sheared marlstone crosscut by boudinage calcite veinlet (drill core HA 486-D; 78.74-78.93 m).

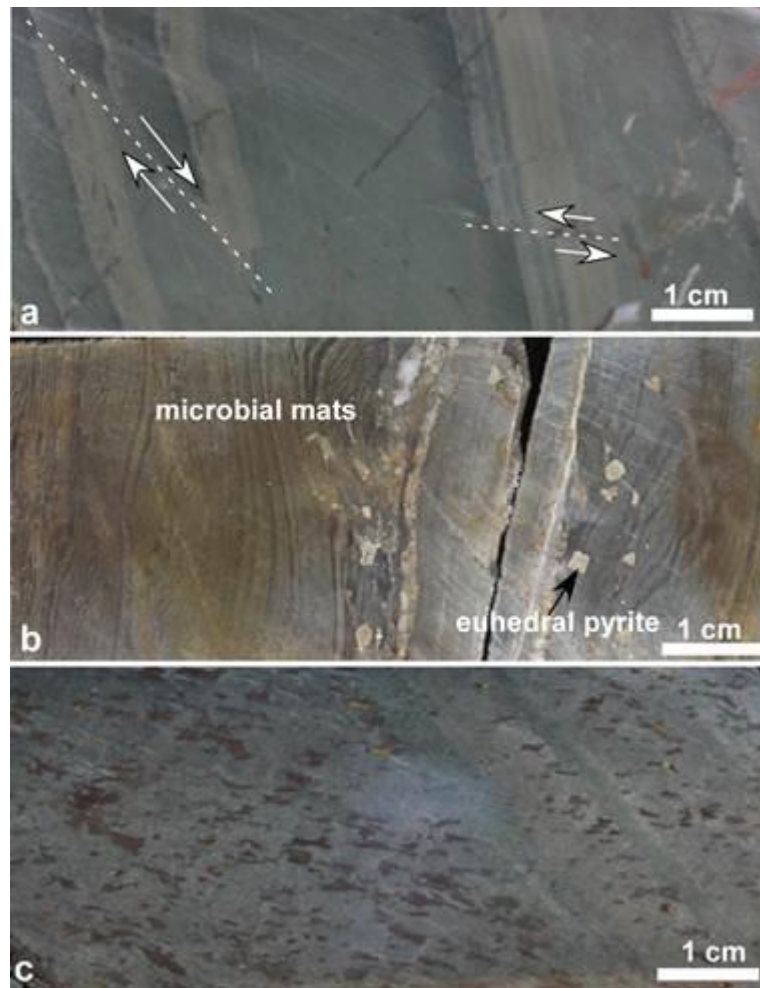


Fig. 3.6: Photographs of mudrocks of the D'Kar Formation illustrating different structures observed in the rocks. (a) Laminated siltstone that has been disturbed by syn-sedimentary normal and thrust faults (drill core HA 520-D; 160.21-160.32 m). (b) Finely laminated siltstone displaying microbial activity that is associated with syn-sedimentary pyrite (drill core HA 10-D; 153.53-153.58 m). (c) Chalcopyrite in nodules and lenses along the bedding (drill core HA 486-D; 78.64-78.74 m)

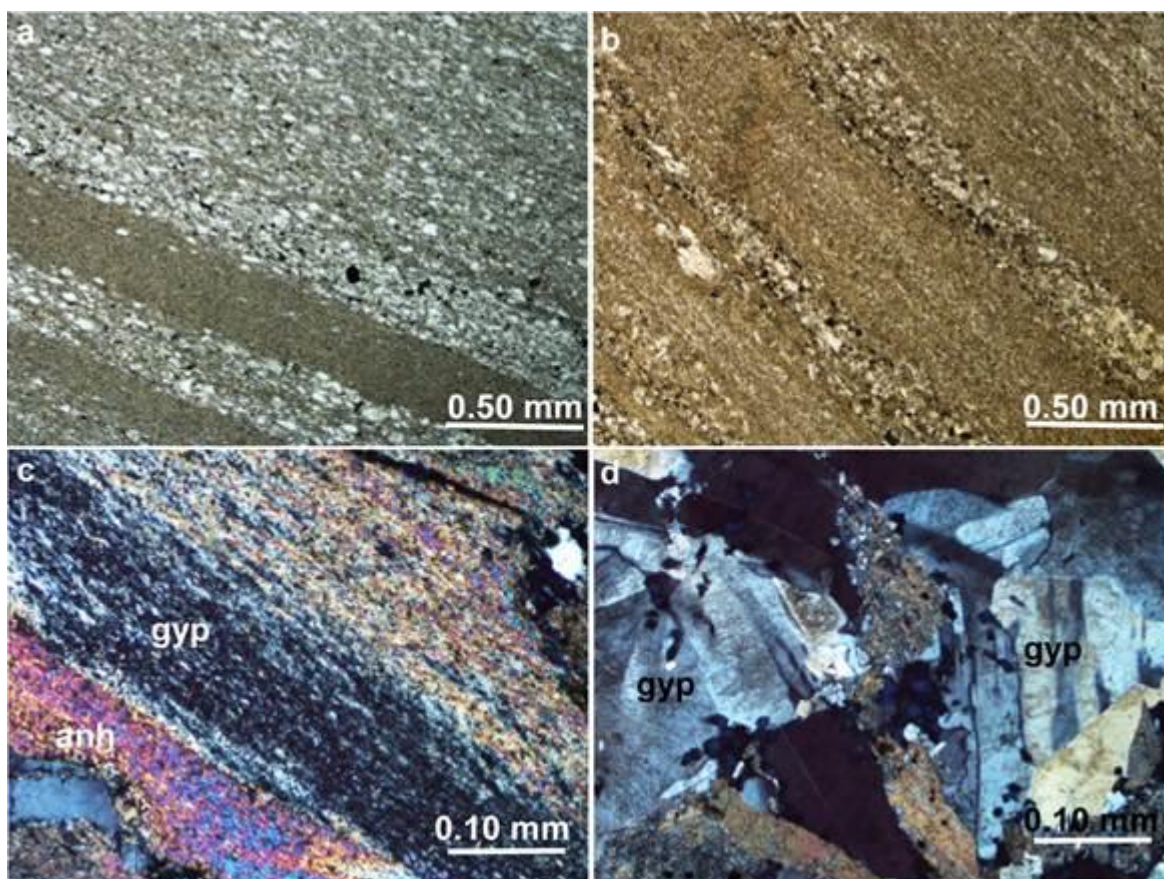


Fig. 3.7: Photomicrographs (transmitted light) of Mudrocks of the D'Kar Formation. (a) Finely laminated siltstone. (b) Finely laminated mudstone. (c) Evaporite beds (anh = anhydrite; gyp = gypsum). (d) Gypsum (gyp) in a brecciated rock. Note the relics of the host rock.

The D'Kar Formation is characterised by fine-grained, massive grey sandstone (Fig. 3.8a, b). Occasionally, the sandstone is oxidised to a reddish brown to pinkish colour (Fig. 3.8c, d). Sandstone of the D'Kar Formation is moderately- to well-sorted (grains size ranging between 0.08 and 0.75 mm) and texturally submature to mature as demonstrated by the absence of clay matrix and moderate- to well-sorting of the grains. This sandstone is characterised by grains of variable roundness similar to the Ngwako Pan Formation sandstone. A few coarse-grained, rounded to well-rounded and clear quartz grains with pitted surface are also observed. The sandstone from the D'Kar Formation has an average compositional ratio of $Qt_{67}: F_{30}: L_3$ (Table 3.1) and is classified as subarkose, and arkose (Fig. 3.2), thus pointing to its compositional immaturity.

Monocrystalline inclusion-free quartz with non-undulose extinction (Fig. 3.3a) is the most common quartz type ranging from 48.9 to 77.6 vol. % (average 64.4 vol.%, Table 3.1).

Few grains with undulose extinction are also observed. Polycrystalline quartz occurs in a relatively small amount (average 2.3 vol. %, Table 3.1). K-feldspar clasts (average 17.4 vol. %) including microcline, orthoclase, and perthite (Figs. 3.3c, e, f), dominate over plagioclase (average 12.9 vol. %) (Fig. 3.3a). Feldspar has undergone variable alteration types, including, argillisation, sericitisation, and carbonatisation. Lithic fragments range from 0.0 to 10.2 vol. % and are largely composed of mudstone (Table 3.1). Metamorphic rock fragments and granitic rock fragments are subordinate (Table 3.1, Fig. 3.4).

The matrix proportion of D'Kar sandstone is highly variable (1.0-12.0 vol. %) and is composed of argillaceous materials (detrital clay, sericite, and chlorite). The grains are cemented by sparry calcite and quartz overgrowths in optical continuity with quartz grains.

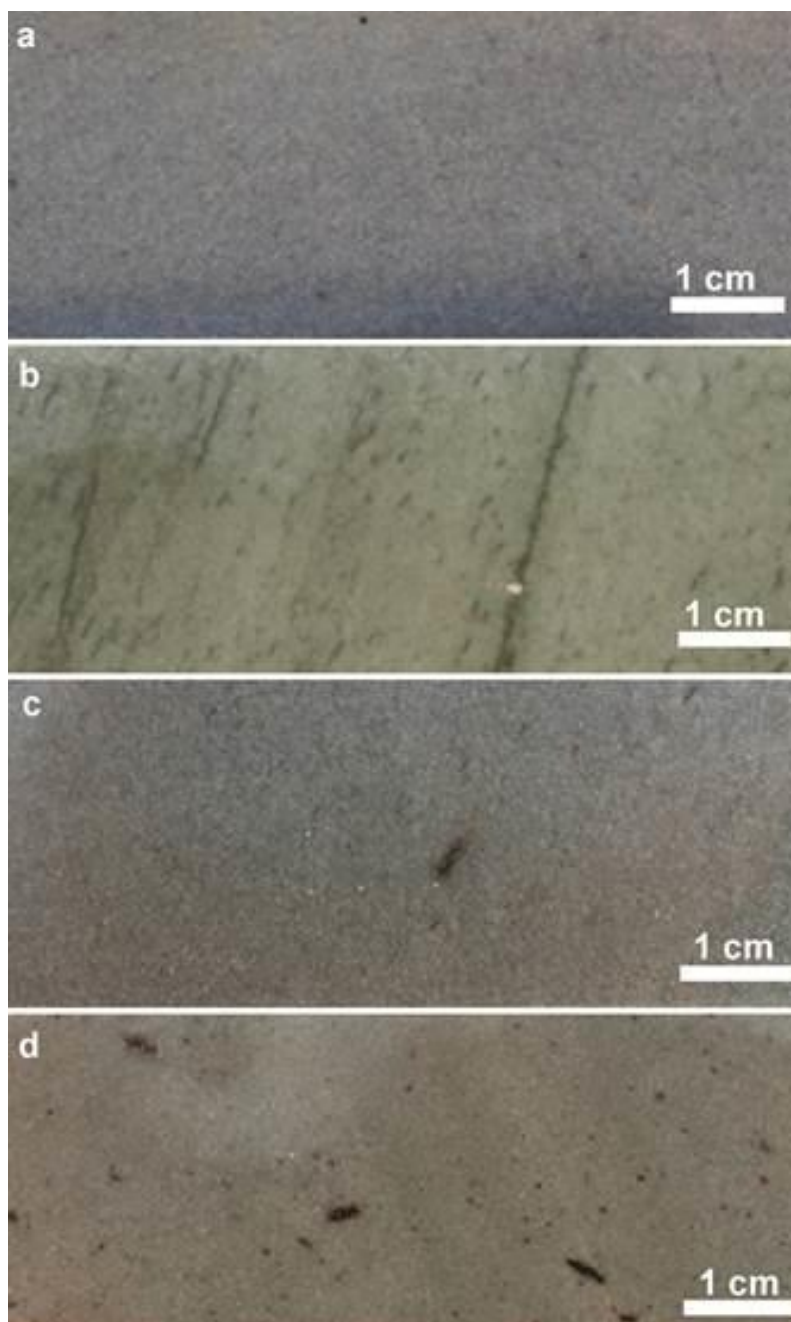


Fig. 3.8: Photographs of the sandstone of the D'Kar Formation showing variation in colour. (a) Brownish grey massive sandstone (drill core HA 515-D; 121.47-122.12 m). (b) Grey and cleaved sandstone (drill core HA 17-D; 206.45-206.64 m). (c) Reddish brown massive sandstone (drill core HA 17-D; 136.80-136.89 m). (d) Pinkish massive sandstone with blebs of hematite (drill core HA 17-D; 192.59-192.69 m).

CHAPTER 4

GEOCHEMICAL COMPOSITIONS OF THE NGWAKO PAN AND THE D’KAR FORMATIONS

This chapter highlights the geochemical characteristics of the sedimentary rocks spatially associated with the Banana Zone Cu-Ag mineralisation.

4.1. Major elements

Major element oxides concentrations are given in Table 4.1 along with major oxide ratios.

4.1.1. Ngwako Pan Formation

Sandstone of the Ngwako Pan Formation has a wide range in SiO_2 contents varying from 64.00 to 75.90 wt.% and a restricted range in Al_2O_3 contents varying from 7.03 to 9.93 wt.%. This sandstone is characterised by TiO_2 contents of 0.20-0.39 wt.%, K_2O contents of 2.98-3.42 wt.% and Na_2O contents of 1.76-2.57 wt.%. The Fe_2O_3 , MgO and CaO contents vary from 0.91 to 1.92 wt.%, 0.35 to 0.72 wt.%, and 2.84 to 6.96 wt.%, respectively. The $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios for the Ngwako Pan Formation sandstone range from 7.90 to 9.89 and from 24.46 to 37.05, respectively. The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ values are between 0.51 and 0.77.

4.1.2. D’Kar Formation

Sandstone of the D’Kar Formation has a restricted range of SiO_2 and Al_2O_3 contents varying from 73.32 to 77.18 wt.% and from 7.41 to 9.54 wt.%, respectively. This

sandstone shows slightly higher contents of TiO_2 , K_2O , and Na_2O (0.24-0.46 wt.%, 2.82-3.50 wt.% and 1.76-2.97 wt.%, respectively) with respect to sandstone of the Ngwako Pan Formation. The Fe_2O_3 and MgO contents vary from 0.92 to 2.38 wt.% and from 0.48 to 0.69 wt.%, respectively. The CaO contents in the D'Kar Formation sandstone samples vary from 2.51 to 5.07 wt.%. The $\text{Al}_2\text{O}_3/\text{TiO}_2$ and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios in these Sandstone samples range from 21.36 to 29.29 and from 0.62 to 1.02, respectively. The $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios range from 7.60 to 9.10.

Mudrocks of the D'Kar Formation are characterised by lower SiO_2 (60.48-64.15 wt.%), but relatively higher Al_2O_3 (12.6-17.67 wt.%) and TiO_2 (0.65-0.86 wt.%) contents with respect to the D'Kar Formation sandstone. The K_2O and Na_2O contents vary from 1.64 to 4.45 wt.% and from 1.90 to 2.84 wt.%, respectively. Mudrocks of the D'Kar Formation are richer in Fe_2O_3 (5.1-6.63 wt.%) and MgO (2.64-3.86 wt.%) in comparison to the sandstone of the D'Kar Formation. The CaO contents range from 1.46 to 5.37 wt.%. The $\text{SiO}_2/\text{Al}_2\text{O}_3$, $\text{Al}_2\text{O}_3/\text{TiO}_2$ and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios range from 3.42 to 5.19, from 18.55 to 21.32 and from 0.44 to 1.74, respectively.

The carbonates (i.e., limestone and marlstone) of the D'Kar Formation show a low content of SiO_2 (14.78-48.88 wt.%) and Al_2O_3 (2.47-8.74 wt.%). TiO_2 content is also low (0.15-0.47 wt.%) compared to the mudrocks and sandstone of the D'Kar Formation. The K_2O and Na_2O contents vary between 0.44 and 1.31 wt.% for K_2O and 0.46 and 2.82 wt.% for Na_2O . The Fe_2O_3 content varies between 1.20 and 4.69 wt.%, which is higher relative to the Fe_2O_3 content of sandstone. MgO and CaO contents vary between 0.60 to 3.03 wt.% and 22.40 to 44.00 wt. %, respectively. The $\text{SiO}_2/\text{Al}_2\text{O}_3$, $\text{Al}_2\text{O}_3/\text{TiO}_2$, and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios range from 5.16 to 5.98, 16.47 to 19.76 and 1.05 to 2.76, respectively.

4.2. Trace elements

Trace element contents of the investigated samples are shown in Table 4.1. The next paragraphs present the distribution of selected trace elements in the Ngwako and D’Kar Formations.

4.2.1. Ngwako Pan Formation

The Cr, Co, and Ni contents in the sandstone of the Ngwako Pan Formation range from 22 to 38 ppm, 6 to 10 ppm and 11 to 22 ppm, respectively. The Th and Zr contents range from 4 to 12 ppm and from 86 ppm to 183 ppm, respectively. The concentration of V in this sandstone is between 29 and 54 ppm. The Ngwako Pan Formation sandstone is relatively rich in large ion lithophile elements (LILE; Ba, Rb and Sr) that vary from 731 to 788 ppm, 84 to 108, and 41 to 61 ppm, respectively (Table 4.1). The La/Co and La/Cr values range from 1.02 to 5.84 and 0.46 to 1.77, respectively. The Th/Cr and Th/Co ratios in the Ngwako Pan Formation sandstone range from 0.18 to 0.52, respectively (Table 4.2). Furthermore, this sandstone has an average Cr/Th ratio of 3.76 (Table 4.2).

4.2.2. D’Kar Formation

The Cr, Co, and Ni contents in the sandstone of the D’Kar Formation range from 15 to 28 ppm, from 3 to 8 ppm and from 9 to 17 ppm, respectively. Thorium contents range from 4 to 5 ppm and Zr contents range from 117 to 181 ppm. The V content varies from 33 to 55 ppm. Barium concentration varies from 733 to 822 ppm, Rb content changes from 66 to 75 ppm, whereas the Sr concentration varies from 48 to 59 ppm (Table 4.1). The La/Co and La/Cr ratios range from 2.82 to 5.24 and 0.79 to 1.17, respectively. The Th/Cr and Th/Co values in this sandstone vary from 0.19 to 0.30 and from 0.61 to 1.34, respectively (Table 4.2). The Cr/Th ratio has an average value of 4.77 (Table 4.2).

In comparison with the sandstone samples, the mudrocks samples of the D’Kar Formation show higher content of Cr, Co and Ni (58-72 ppm, 15-38 ppm and 23-33 ppm, respectively). Thorium contents in the mudrocks vary from 6 to 11 ppm. D’Kar Formation mudrocks are also characterised by relatively high Zr and V contents (153 ppm to 236 ppm and 98 to 130 ppm, respectively). The mudrocks of the D’Kar Formation are rich in Ba, Rb and Sr ranging from 323 to 945 ppm, 51 to 264 ppm and 35 to 64 ppm, respectively (Table 4.1). The La/Co and La/Cr values range from 0.80 to 2.62 and 0.27 to 0.70, respectively, while the Th/Cr and Th/Co ratios vary from 0.09 to 0.18 and 0.25 to 0.69, respectively (Table 4.2). The mudrocks samples are characterised by high Cr/Th ratio with an average value of 7.31 (Table 4.2).

In comparison to the sandstone and mudrocks of the D’kar Formation, the carbonate rocks of D’Kar Formation are also enriched in transition elements (Table 4.1) especially Co and V, which vary from 4 to 42 ppm and 23 to 86 ppm, respectively. However, the carbonate rocks higher concentration of Sr (155-656 ppm) and lower concentration of Rb (24-46 ppm) relative to the sandstone and mudrocks. Barium content in the carbonates varies from 161 to 938 ppm, and Ba concentration show a positive correlation with the Sr concentration and a negative correlation with Rb.

Table 4.1

Major, minor, and trace element concentrations, major oxide ratios and Chemical Index of Weathering (CIW') of the Ngwako Pan Formation sandstone, D'Kar Formation sandstone, D'Kar Formation mudrocks and D'Kar Formation carbonate rocks.

	Ngwako Pan Formation sandstone						D'Kar Formation sandstone						D'Kar Formation mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg	MS1	MS2	MS3	MS4	MS5	Avg	MD1	MD2	MD3	MD4	MD5	Avg
Major and minor elements (wt. %)																		
SiO ₂	73.3	75.3	77.2	76.0	76.6	75.7	64.0	75.5	74.7	75.9	75.7	73.2	63.9	64.2	61.7	61.9	60.5	62.4
TiO ₂	0.20	0.39	0.29	0.30	0.29	0.29	0.24	0.46	0.35	0.42	0.45	0.38	0.79	0.75	0.65	0.73	0.86	0.76
Al ₂ O ₃	7.41	9.54	8.43	8.55	8.49	8.48	7.03	9.93	9.48	9.61	9.77	9.17	15.9	14.5	12.	15.6	17.7	15.1
Fe ₂ O ₃ (t)	0.91	1.92	1.53	1.47	1.50	1.46	0.92	2.38	1.65	2.16	2.27	1.88	5.10	5.55	5.47	6.63	5.34	5.62
MnO	0.16	0.09	0.09	0.11	0.10	0.11	0.10	0.08	0.09	0.10	0.09	0.09	0.11	0.14	0.24	0.12	0.11	0.14
MgO	0.35	0.72	0.41	0.50	0.45	0.49	0.48	0.69	0.59	0.52	0.60	0.57	2.82	2.64	3.86	2.96	3.33	3.12
CaO	6.96	2.84	3.66	4.55	4.10	4.42	5.07	2.51	3.79	3.64	3.08	3.62	1.54	2.46	5.37	2.48	1.46	2.66
Na ₂ O	1.76	2.57	2.29	2.23	2.26	2.22	1.76	2.97	2.37	2.26	2.62	2.39	1.96	2.8	2.84	1.9	2.84	2.47
K ₂ O	3.42	3.35	2.98	3.29	3.13	3.23	2.82	2.92	3.50	2.85	2.88	2.99	4.45	3.02	1.63	3.71	4.35	3.43
P ₂ O ₅	0.06	0.09	0.07	0.07	0.07	0.07	0.06	0.11	0.09	0.07	0.09	0.08	0.19	0.18	0.17	0.18	0.21	0.19
Cr ₂ O ₃	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01
LOI	5.03	2.21	2.76	2.38	2.27	2.93	16.74	2.16	2.82	2.31	2.24	5.25	2.8	3.35	5.65	3.42	2.93	3.63
Total	99.6	99.1	99.7	99.4	99.3	99.4	99.2	99.7	99.5	99.9	99.8	99.6	99.5	99.5	99.7	99.6	99.6	99.6
Na₂O/K₂O	0.51	0.77	0.77	0.68	0.72	0.69	0.62	1.02	0.68	0.79	0.91	0.80	0.44	0.93	1.74	0.51	0.65	0.85
SiO₂/Al₂O₃	9.89	7.90	9.16	8.89	9.02	8.97	9.10	7.60	7.88	7.90	7.74	8.05	4.02	4.43	5.12	3.97	3.42	4.19
Al₂O₃/TiO₂	37.1	24.5	29.1	28.8	29.0	29.7	29.3	21.6	27.1	21.4	23.3	24.5	20.1	19.3	18.6	21.3	20.6	20.0
CIW'	81.0	79.0	79.0	79.0	79.0	79.0	80.0	77.0	80.0	81.0	79.0	79.0	89.0	84.0	81.0	89.0	86.0	86.0

Table 4.1 (continued).

	D'kar Formation carbonate rocks					
	CD1	CD2	CD3	CD4	CD5	Avg
SiO₂	40.6	40.9	48.9	42.8	14.8	37.6
TiO₂	0.40	0.42	0.47	0.38	0.15	0.36
Al₂O₃	7.87	7.68	8.74	7.51	2.47	6.85
Fe₂O₃ (t)	3.72	2.89	4.69	3.40	1.20	3.18
MnO	0.59	0.49	0.43	0.49	0.49	0.50
MgO	1.52	2.26	3.03	2.18	0.60	1.92
CaO	24.3	23.4	25.4	22.4	44.0	27.9
Na₂O	2.82	2.19	1.79	1.88	0.46	1.83
K₂O	1.02	0.94	1.31	1.15	0.44	0.97
P₂O₅	0.13	0.13	0.14	0.12	0.11	0.13
Cr₂O₃	0.01	0.01	0.01	0.01	0.01	0.01
LOI	15.9	17.9	4.53	16.9	34.2	17.9
Total	98.9	99.2	99.4	99.2	98.9	99.1
Na₂O/K₂O	2.76	2.33	1.37	1.63	1.05	1.83
SiO₂/Al₂O₃	5.16	5.33	5.59	5.70	5.98	5.55
Al₂O₃/TiO₂	19.7	18.3	18.6	19.8	16.5	18.6
CIW'	-	-	-	-	-	-

Table 4.1 (continued).

	Ngwako Pan Formation sandstone						D’Kar Formation sandstone						D’Kar Formation mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg	MS1	MS2	MS3	MS4	MS5	Avg	MD1	MD2	MD3	MD4	MD5	Avg
Trace elements (ppm)																		
Li	6.04	19.8	14.5	13.6	14.0	13.6	6.22	12.8	9.49	10.3	11.5	10.1	37.4	36.3	28.3	50.1	46.7	39.8
Be	0.79	1.30	1.56	1.22	1.39	1.25	1.16	0.66	0.91	1.08	0.87	0.94	4.10	2.83	1.58	3.70	4.49	3.34
V	29.3	54.0	39.7	41.4	40.6	41.0	33.8	550	44.4	47.3	51.1	46.3	116	106	98.1	119	130	114
Cr	22.5	37.7	22.3	27.9	25.1	27.1	15.5	27.6	21.5	22.5	25.0	22.4	71.3	64.2	58.4	64.0	72.1	66.0
Co	10.1	8.78	6.77	8.65	7.71	8.39	3.45	7.72	5.59	6.16	6.94	5.97	24.0	15.9	37.6	16.4	36.5	26.1
Ni	11.9	19.8	22.2	18.1	20.1	18.4	9.00	16.5	12.8	13.3	14.9	13.3	27.6	29.4	32.5	23.6	24.2	27.4
Cu	4108	1872	2500	2861	2680	2804	1005	37.6	521	843	440	569	5705	4205	2274	648	4914	3550
Zn	22.9	82.9	64.8	57.2	61.0	57.8	22.1	776	399	148	462	361	121	162	238	96.9	175	159
Ga	7.44	12.2	10.6	10.2	10.4	10.2	8.13	11.8	9.98	8.75	10.3	9.80	27.2	23.4	18.8	25.4	29.0	24.8
Rb	84.7	108	97.8	97.7	97.7	97.1	75.4	66.2	70.8	73.8	70.0	71.2	264	120	51.9	177	206	164
Sr	60.9	41.7	54.7	52.9	53.8	52.8	58.5	48.6	53.5	56.8	52.7	54.0	35.7	58.4	63.8	39.6	38.7	47.2
Mo	1.15	51.9	2.73	19.1	10.9	17.2	12.6	0.49	6.56	10.6	5.54	7.16	8.12	2.37	29.1	5.77	3.88	9.84
Ag	< 0.50	< 0.50	< 0.50	< 0.50	< 0.50	-	< 0.50	< 0.50	< 0.50	< 0.50	< 0.50	-	< 0.50	< 0.50	< 0.50	< 0.50	< 0.50	-
Cd	0.18	0.24	0.14	0.19	0.16	0.18	0.16	8.01	4.09	1.47	4.74	3.69	0.08	0.10	0.26	0.14	0.21	0.16
Te	< 0.01	< 0.01	< 0.01	< 0.01	< 0.01	-	< 0.01	< 0.01	< 0.01	< 0.01	< 0.01	-	< 0.01	0.02	0.02	0.01	< 0.01	0.02
Ba	788	750	731	764	748	756	733	822	778	748	785	774	822	610	323	614	945	663
Tl	0.46	0.54	0.56	0.52	0.54	0.53	0.41	0.39	0.40	0.41	0.40	0.40	1.05	0.54	0.27	0.83	1.21	0.78
Pb	43.6	9.50	14.8	23.0	18.9	22.0	5.56	513	259	90.1	302	234	4.16	6.90	7.70	72.6	4.39	19.2
Bi	67.6	2.95	3.77	25.4	14.6	22.9	1.42	0.16	0.79	1.21	0.69	0.85	4.14	0.41	2.85	0.54	1.31	1.85
U	2.50	2.03	1.35	1.99	1.67	1.91	1.77	1.34	1.56	1.70	1.52	1.58	3.78	3.98	6.75	4.07	4.84	4.68
Th	4.40	6.85	11.6	7.63	9.63	8.03	4.63	4.70	4.67	4.84	4.65	4.70	6.18	9.70	9.32	11.4	11.0	9.53
Zr	86.2	168	183	147	165	150	117	181	150	148	144	148	154	218	236	186	190	197

Table 4.1 (continued and end).

	D’Kar Formation carbonate rocks					
	CD1	CD2	CD3	CD4	CD5	Avg
Li	11.5	14.2	16.7	15.1	6.33	12.8
Be	0.76	1.14	1.16	0.66	0.68	0.88
V	57.6	69.4	79.2	86.2	23.9	63.3
Cr	33.8	34.4	34.2	31.8	11.5	29.1
Co	17.1	17.5	42.3	17.7	4.31	19.8
Ni	19.1	17.3	15.0	20.7	13.7	17.2
Cu	26095	10879	13220	17688	52.7	13587
Zn	104	84.2	83.3	129	91.2	98.4
Ga	9.70	11.3	11.0	9.94	3.40	9.06
Rb	43.9	46.0	39.0	30.4	24.7	36.8
Sr	216	222	214	115	656	285
Mo	1.68	40.5	65.0	86.7	1.51	39.1
Ag	< 0.50	< 0.50	< 0.50	< 0.50	< 0.50	-
Cd	0.74	0.21	0.55	0.79	0.44	0.55
Te	< 0.01	< 0.01	< 0.01	< 0.01	< 0.01	-
Ba	164	161	234	231	938	346
Tl	0.39	0.25	0.21	0.17	0.14	0.23
Pb	28.1	5.15	24.1	233	6.89	59.4
Bi	6.91	9.34	18.2	19.2	0.19	10.8
U	5.06	5.44	4.76	13.2	10.3	7.75
Th	5.32	6.04	6.54	6.34	2.91	5.43
Zr	19.0	23.1	24.9	22.3	10.7	20.0

4.3. Rare earth elements

The REE data are presented in Table 4.2 and are shown as chondrite-normalised patterns in Fig. 4.1a, b for the sandstone of the Ngwako Pan and the D’Kar Formations and as PAAS-normalised patterns in Fig. 4.1c, d for the mudrocks and carbonate rocks, respectively. The total REE contents (ΣREE) of the investigated samples range from 56.0 to 211 ppm (average 138 ppm) in the sandstone of the Ngwako Pan Formation, from 91.0 to 123 ppm (average 111 ppm) in the sandstone of the D’Kar Formation and from 105 to 211 ppm (average 188 ppm) in the mudrocks of the D’Kar Formation. The carbonate rocks of the D’Kar Formation have a wide range from 47.0 to 125 ppm (average 96 ppm).

4.3.1. Ngwako Pan Formation

When normalised to chondrite, the sandstone of the Ngwako Pan Formation (Fig. 4.1a) reveals a high LREE/HREE ratios (5.04-6.90; average 6.16), an enriched LREE ($\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}} = 3.27\text{-}2.68$; average 2.95 ± 0.21) and a flat HREE ($\text{Gd}_{\text{CN}}/\text{Yb}_{\text{CN}} = 1.41\text{-}1.09$; average 1.30 ± 0.13) pattern that is very similar to PAAS except for low $\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}}$ ratios. These samples show a negative to slightly negative Eu anomaly ($\text{Eu}/\text{Eu}^* = 0.61\text{-}0.83$; average 0.70 ± 0.09) and a slightly negative Ce anomaly ($\text{Ce}/\text{Ce}^* = 0.91\text{-}0.93$; average 0.92 ± 0.01). The Y/Ho ratios vary from 26.9 to 27.6 (average 27.1).

4.3.2. D’Kar Formation

The chondrite normalised REE pattern for the sandstone of the D’Kar Formation is presented in Fig. 4.1b. All the samples show a lower LREE contents relative to PAAS, whereas HREE values are very similar to PAAS. The samples are characterised by high LREE/HREE ratios (5.71-6.55; average 6.27), LREE enrichment ($\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}} = 3.61\text{-}2.68$; average 3.13 ± 0.42) coupled with flat HREE profile ($\text{Gd}_{\text{CN}}/\text{Yb}_{\text{CN}} = 1.30\text{-}1.11$; average

1.19±0.07) and a negative to slightly negative Eu anomaly ($\text{Eu}/\text{Eu}^*=0.69\text{--}0.81$; average 0.75 ± 0.05). The Ce anomaly is slightly negative ($\text{Ce}/\text{Ce}^*=0.83\text{--}0.91$; average 0.90 ± 0.04), and Y/Ho ratios between 28.3 and 31.3 (average 29.5).

All, but one sample (MD1) of the mudrocks of the D'Kar Formation show an overall flat PAAS-normalised REE pattern (average $\text{Pr}_{\text{SN}}/\text{Yb}_{\text{SN}}=0.84\pm0.11$ without MD1) (Fig. 4.1c). Mudrocks are characterised by variable Eu anomalies ($\text{Eu}/\text{Eu}^*=0.81\text{--}1.21$; average 0.93 ± 0.16), a slightly negative Ce anomaly ($\text{Ce}/\text{Ce}^*=0.86\text{--}0.90$; average 0.87 ± 0.05), and Y/Ho ratios between 26.3 and 28.8 (average 27.9).

The carbonate rocks of the D'Kar Formation show a flat PAAS-normalised REE pattern (average $\text{Pr}_{\text{SN}}/\text{Yb}_{\text{SN}}=0.71\pm0.05$) (Fig. 4.1d). The carbonate rocks are characterised by Eu anomalies that are variable ($\text{Eu}/\text{Eu}^*=0.76\text{--}1.29$; average 0.99 ± 0.21), a slightly negative to no Ce anomalies ($\text{Ce}/\text{Ce}^*=0.86\text{--}1.02$; average 0.94 ± 0.06) and Y/Ho chondrite ratios between 27.8 and 29.7 (average 28.7).

Table 4.2

Rare earth elements and Yttrium (REE+Y) abundance (ppm), Σ REE, Y/Ho, REE ratios and Eu and Ce anomalies of the Ngwako Pan Formation sandstone, D'Kar Formation sandstone, D'Kar Formation mudrocks and D'Kar Formation carbonate rocks.

	Ngwako Pan Formation sandstone						D'Kar Formation sandstone						D'Kar Formation mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg.	MS1	MS2	MS3	MS4	MS5	Avg.	MD1	MD2	MD3	MD4	MD5	Avg.
Y	13.6	21.5	40.9	25.3	33.1	26.9	18.4	25.9	23.2	26.6	25.8	24.0	25.1	36.8	31.9	38.2	40.4	34.5
La	10.3	23.5	39.5	24.4	32.0	25.9	18.1	21.8	20.9	23.7	24.8	21.9	19.2	41.7	33.5	35.6	50.6	36.1
Ce	20.2	47.6	80.5	49.3	64.9	52.5	35.1	42.2	39.7	44.3	45.8	41.4	38.5	81.3	67.4	72.3	102	72.3
Pr	2.66	6.12	10.57	6.43	8.50	6.86	4.48	5.60	6.04	6.17	5.73	5.60	5.01	10.7	8.95	9.42	13.2	9.47
Nd	10.7	23.4	41.7	25.2	33.5	26.9	17.5	21.9	20.7	24.3	24.6	21.8	20.5	41.6	34.8	37.1	51.0	37.0
Sm	2.41	4.52	8.56	5.15	6.85	5.50	3.22	4.60	4.91	5.25	4.33	4.46	4.55	8.09	7.09	7.78	9.68	7.44
Eu	0.62	1.05	1.62	1.10	1.36	1.15	0.78	1.00	1.03	1.08	1.11	1.00	1.14	1.44	1.15	1.42	1.63	1.36
Gd	2.15	4.03	7.74	4.62	6.18	4.95	2.97	4.3	3.64	3.99	4.05	3.79	4.18	7.21	6.16	6.81	8.47	6.57
Tb	0.40	0.71	1.39	0.83	1.11	0.89	0.54	0.79	0.67	0.68	0.74	0.68	0.77	1.24	1.07	1.21	1.45	1.15
Dy	2.43	4.07	7.83	4.76	6.30	5.08	3.13	4.63	3.88	4.38	4.58	4.12	4.52	6.96	5.98	6.84	7.86	6.43
Ho	0.50	0.78	1.52	0.93	1.23	0.99	0.64	0.90	0.77	0.85	0.91	0.81	0.87	1.35	1.14	1.34	1.51	1.24
Er	1.58	2.43	4.37	2.79	3.58	2.95	1.98	2.73	2.36	2.42	2.77	2.45	2.64	3.99	3.51	4.04	4.40	3.72
Tm	0.25	0.38	0.68	0.44	0.56	0.46	0.32	0.41	0.38	0.42	0.41	0.39	0.42	0.63	0.52	0.62	0.69	0.58
Yb	1.60	2.47	4.45	2.84	3.64	3.00	2.06	2.68	2.66	2.71	2.77	2.58	2.65	4.11	3.46	3.97	4.46	3.73
Lu	0.27	0.37	0.68	0.44	0.56	0.46	0.33	0.39	0.38	0.43	0.42	0.39	0.42	0.65	0.54	0.61	0.65	0.57
ΣREE	56.1	121	211	129	170	138	91.2	114	108	121	123	111	105	211	175	189	257	188
(ΣLREE)	46.3	105	181	110	146	118	78.4	96.1	92.3	104	105	95.1	87.7	183	152	162	226	162
(ΣHREE)	9.18	15.2	28.7	17.7	23.2	18.8	12.0	16.8	14.7	15.9	16.7	15.2	16.5	26.1	22.4	25.4	29.5	24.0
LREE/HREE	5.04	6.90	6.31	6.26	6.29	6.16	6.55	5.71	6.26	6.53	6.32	6.27	5.33	7.01	6.78	6.37	7.67	6.63
(Eu/Eu*)_{CN}	0.83	0.75	0.61	0.69	0.64	0.70	0.77	0.69	0.74	0.72	0.81	0.75	-	-	-	-	-	-
(Eu/Eu*)_{SN}	-	-	-	-	-	-	-	-	-	-	-	-	1.21	0.88	0.81	0.90	0.84	0.93
(Ce/Ce*)_{CN}	0.91	0.93	0.92	0.92	0.92	0.92	0.91	0.90	0.83	0.86	0.90	0.90	-	-	-	-	-	-
(Ce/Ce*)_{SN}	-	-	-	-	-	-	-	-	-	-	-	-	0.97	0.86	0.86	0.90	0.86	0.89
(La/La*)_{SN}	-	-	-	-	-	-	-	-	-	-	-	-	1.00	0.92	0.89	0.92	0.89	0.92

Table 4.2 (continued).

La_{CN}/Sm_{CN}	2.68	3.27	2.91	2.98	2.94	2.95	3.53	2.98	2.68	2.84	3.61	3.13	-	-	-	-	-	-
Gd_{CN}/Yb_{CN}	1.09	1.32	1.41	1.32	1.38	1.30	1.17	1.30	1.11	1.19	1.18	1.19	-	-	-	-	-	-
(Pr/Yb)_{SN}	-	-	-	-	-	-	-	-	-	-	-	-	0.60	0.83	0.83	0.76	0.95	0.79
Y/Ho	27.1	27.6	26.9	27.2	27.0	27.1	28.7	28.8	30.1	31.3	28.3	29.5	28.8	27.2	27.9	28.5	26.8	27.9

(LREE/HREE) = $\Sigma(\text{La-Sm}) / \Sigma(\text{Gd-Lu})$; Subscript _{CN} refers to chondrite-normalised values; Subscript _{SN} refers to PAAS-normalised values.

Table 4.2 (continued).

D'Kar Formation carbonate rocks						
	CD1	CD2	CD3	CD4	CD5	Avg.
Y	19.0	23.1	24.9	22.3	10.7	20.0
La	18.7	22.3	22.0	18.0	8.59	17.9
Ce	38.8	43.0	51.5	36.0	17.7	37.4
Pr	4.78	5.77	5.92	4.69	2.26	4.68
Nd	18.5	22.6	23.4	18.7	9.19	18.5
Sm	3.70	4.57	4.74	4.09	1.91	3.80
Eu	0.84	0.71	0.89	0.74	0.50	0.74
Gd	3.33	4.16	4.30	3.80	1.69	3.46
Tb	0.60	0.73	0.76	0.67	0.30	0.61
Dy	3.44	4.23	4.26	3.93	1.79	3.53
Ho	0.68	0.83	0.85	0.78	0.36	0.70

Table 4.2 (continued and end).

Er	2.08	2.51	2.59	2.35	1.05	2.12
Tm	0.31	0.40	0.40	0.35	0.16	0.32
Yb	2.01	2.58	2.60	2.35	1.05	2.12
Lu	0.32	0.40	0.41	0.35	0.17	0.33
ΣREE	98.0	114.7	125	96.8	46.7	96.2
(ΣLREE)	84.4	98.2	108	81.4	39.6	82.3
(ΣHREE)	12.8	15.8	16.2	14.6	6.62	13.2
LREE/HREE	6.61	6.20	6.65	5.59	5.98	6.21
(Eu/Eu*)_{CN}	-	-	-	-	-	-
(Eu/Eu*)_{SN}	1.11	0.76	0.92	0.88	1.29	0.99-
(Ce/Ce*)_{CN}	-	-	-	-	-	-
(Ce/Ce*)_{SN}	0.91	0.86	1.02	0.92	0.97	0.94
(La/La*)_{SN}	-	-	-	-	-	-
La_{CN}/Sm_{CN}	-	-	-	-	-	-
Gd_{CN}/Yb_{CN}	-	-	-	-	-	-
(Pr/Yb)_{SN}	0.76	0.71	0.73	0.64	0.69	0.71
Y/Ho	27.9	27.8	29.3	28.6	29.7	28.7

(LREE/HREE) = $\Sigma(\text{La-Sm}) / \Sigma(\text{Gd-Lu})$; Subscript _{CN} refers to chondrite-normalised values; Subscript _{SN} refers to PAAS-normalised values.

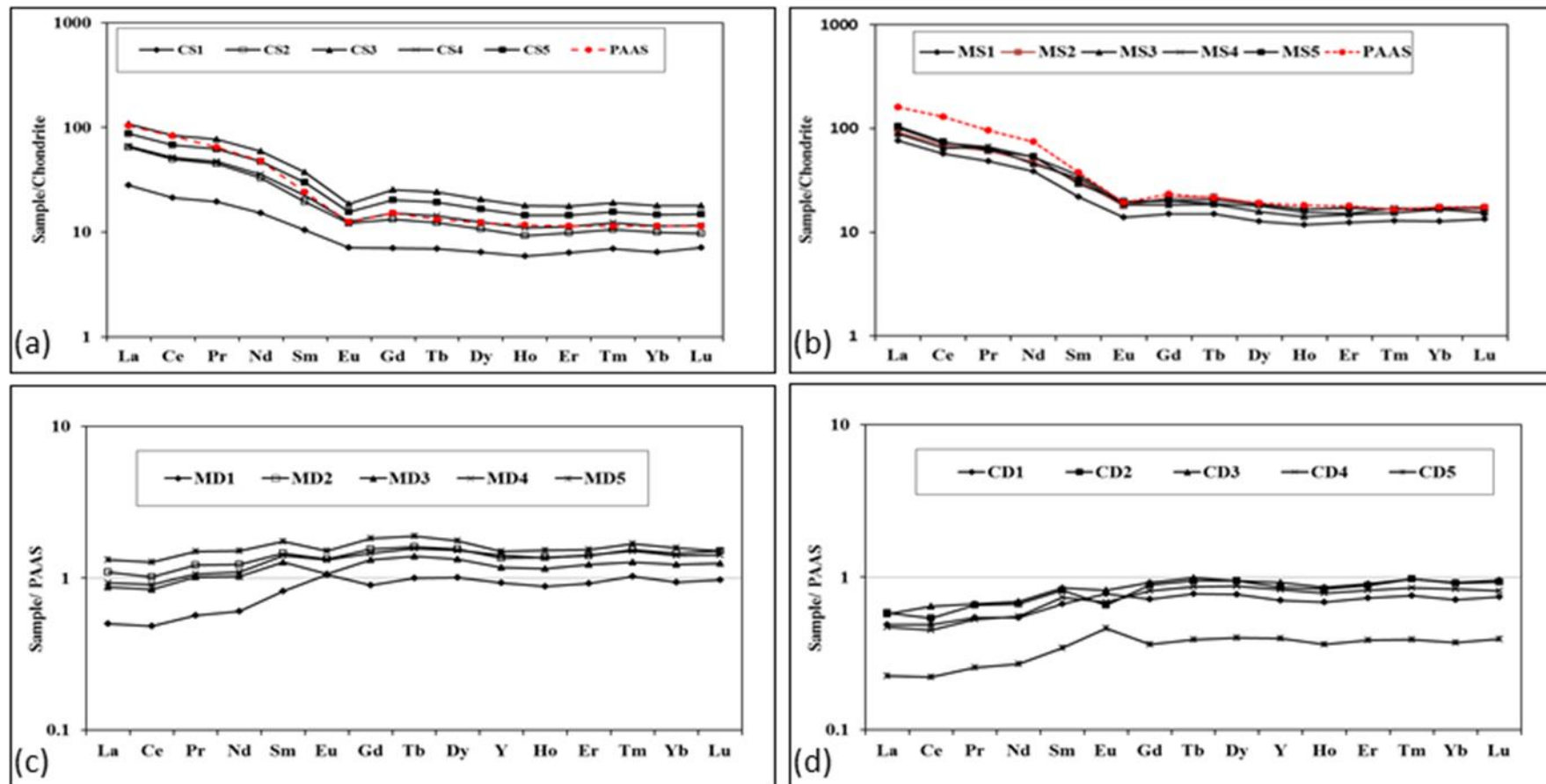


Fig. 4.1: Normalised REE patterns of the studied sediments. (a) Chondrite-normalised Ngwako Pan Formation sandstone. (b). Chondrite-normalised D'Kar Formation sandstone. (c) PAAS-normalised D'Kar Formation mudrocks. (d) PAAS-normalised D'Kar Formation carbonates

Table 4.3

Range of elemental ratios of the samples from this study compared to the ratios in similar fractions derived from felsic and mafic rocks.

Elemental ratio	Range of sediments		This study		
	Felsic Rocks ¹	Mafic Rocks ¹	Ngwako Pan Formation sandstone	D’Kar Formation sandstone	D’Kar Formation mudrocks
Eu/Eu*	0.40-0.94	0.71-0.95	0.60-0.82	0.68-0.80	0.81-1.21
Th/Co	0.67-19.4	0.04-1.40	0.44-11.72	0.61-1.34	0.25-0.69
Th/Cr	0.13-2.70	0.02-0.05	0.18-0.52	0.17-0.30	0.09-0.18
Cr/Th	4.00-15.0	25.0-500	1.92-5.50	3.34-5.87	5.63-11.5

¹ After **Cullers and Podkovyrov, 2000**.

CHAPTER 5

MINERALISATION

The Banana Zone Cu-Ag deposit is classified as a sediment-hosted copper deposit, based on **Meyer's** classification (**1981**). The mineralisation is stratigraphically confined to the shallow marine facies (organic matter-rich mudrocks and fine-grained sandstone) in the D'Kar Formation (see Table 2.1 in Chapter 2), thus indicative of a stratabound mineralisation type. These sediment-hosted Cu-Ag deposits are thought to have been deposited under reducing conditions (cf. **Cailteux et al., 2005**).

Within the Banana Zone, two mineralisation styles were identified and these include: (i) disseminated mineralisation; and (ii) structure-controlled mineralisation.

5.1. Disseminated mineralisation

The disseminated mineralisation consists of scattered sulphides of variable proportion (up to 10 vol. % locally, Fig. 5.1a) either randomly distributed in the host rocks or confined along the stratification/bedding planes (Fig. 5.2a), thus supporting its stratiform nature or hosted within pore spaces. The sulphides are very fine (sulphide dust)- to coarse-grained and are of variable shapes including: (i) individual anhedral grains, (ii) clusters with aggregation of numerous grains (Fig. 5.1a) and (iii) porphyroblasts and boudins (Fig. 5.2b). Disseminated mineralisation style in the area is mainly associated with quartz and calcite (after plagioclase) and locally, with iron oxides/hydroxides in the Ngwako Pan Formation. Disseminated sulphides are essentially composed of chalcopyrite, bornite, and chalcocite (chalcocite here refers to chalcocite group minerals; **Ramdohr, 1980**), with minor pyrite (Fig. 5.2e), pyrrhotite (Fig. 5.2e), arsenopyrite (Fig. 5.2f), sphalerite (Fig. 5.2e), and galena. Pyrite forms framboids, mainly in organic-rich shale (Fig. 5.2c). But it also occurs as medium-grained subhedral crystals, and coarse to very coarse-grained euhedral crystals (Fig. 5.1b) consisting of exclusively of framboidal aggregates (Fig. 5.2d). Chalcopyrite usually is found to replace or rims the framboidal pyrite (Fig. 5.2d). Locally, fine-grained anhedral pyrrhotite and chalcopyrite (Fig. 5.2g) clusters are found

disseminated in organic-rich sediments. In addition to the above mentioned, chalcopyrite and bornite occur as intergrowth or as separate anhedral grains or bornite included in chalcopyrite or vice versa. Furthermore, bornite and chalcocite exhibit an exsolution texture implying their concomitant formation (Fig. 5.2h). It must be noted that the grain size of the mineralisation is positively correlated with the grain size of the host rock. Locally, the intergranular micro porosity within the sandstone is filled with fine-grained sulphides (chalcopyrite, bornite, chalcocite, and pyrite) (Fig. 5.2g, i). Replacement of primary sulphides by covellite (Fig. 5.2j) and intergrowth microtextures have been observed.

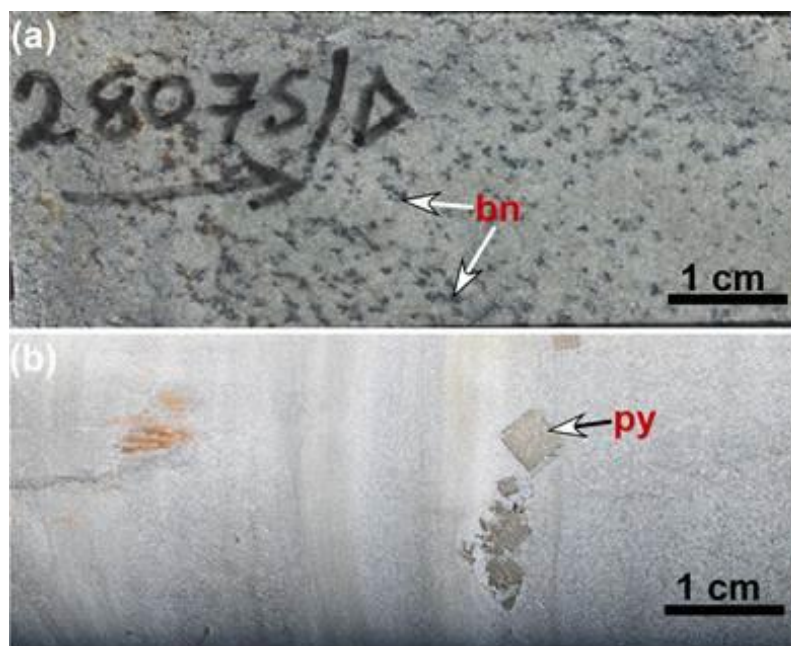


Fig. 5.1: Photographs of drill cores showing the different styles of mineralisation from the Banana Zone. (a) Disseminated bornite (bn) mineralisation (drill core HA 526-D; 145.65-145.75 m). (b) Very coarse euhedral pyrite (py) grains occurring along bedding/stratification planes (drill core HA 248-D; 249.15 m).

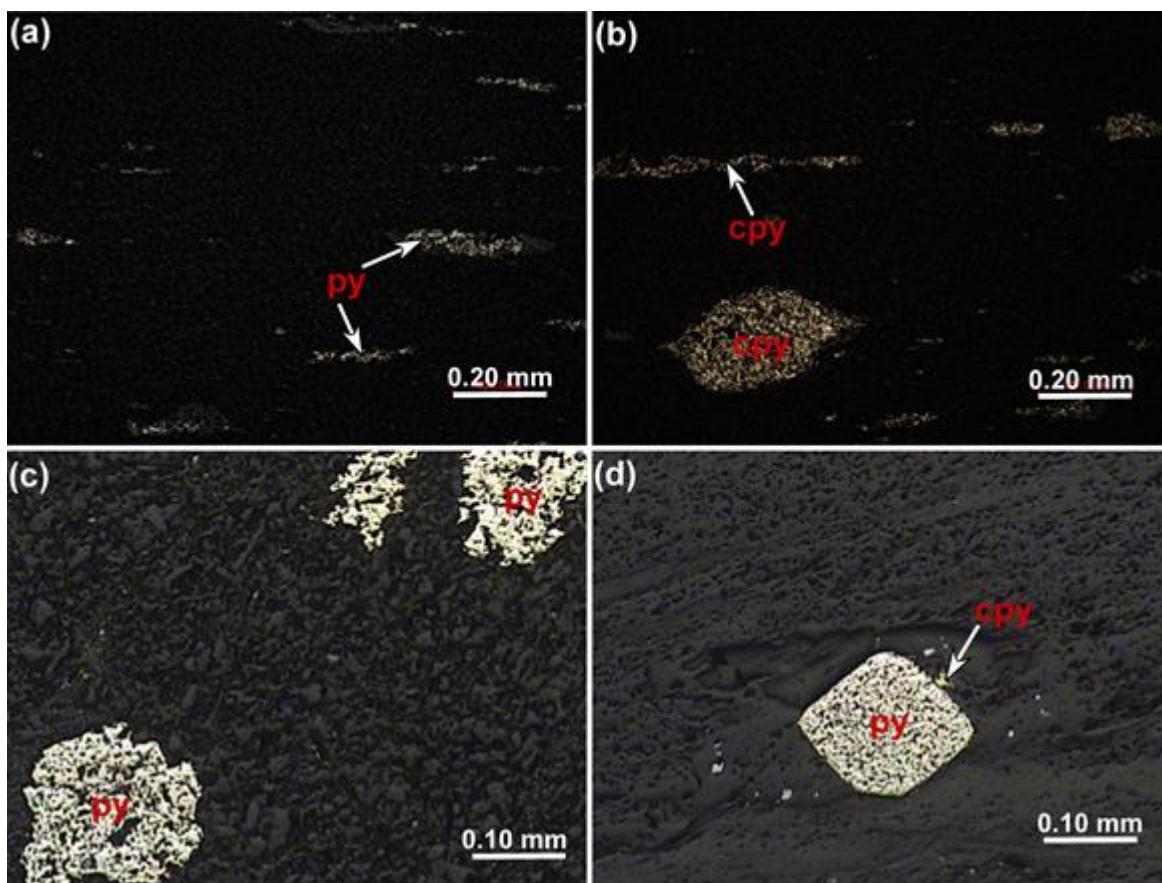


Fig. 5.2: Photomicrographs of polished thin sections (reflected light) showing the types and occurrences of mineralisation of the Banana Zone. (a) Disseminated pyrite (py) in host rock, slightly concentrated along bedding planes. (b) Deformed chalcopyrite (cpy) mineralisation occurring as porphyroblast and boudins. (c) Pyrite (py) displaying a framboidal texture. (d) Euhedral pyrite (py) grain consisting of exclusively of framboidal aggregates. (e) Sphalerite (sph) intergrown with pyrite (py) and pyrrhotite (po). (f) Arsenopyrite (asp) overgrown by bornite (bn). (g) Clusters composed of very fine-grained anhedral pyrrhotite (po) and pitted chalcopyrite (cpy) in a dark/organic-rich rock. (h) Exsolution lamella of chalcopyrite (cpy) in bornite (bn). (i) Micro-pore spaces between quartz grains filled by chalcocite (cc)-bornite (bn) mineralisation. (j) Covellite (cv) completely replacing chalcocite (cc).

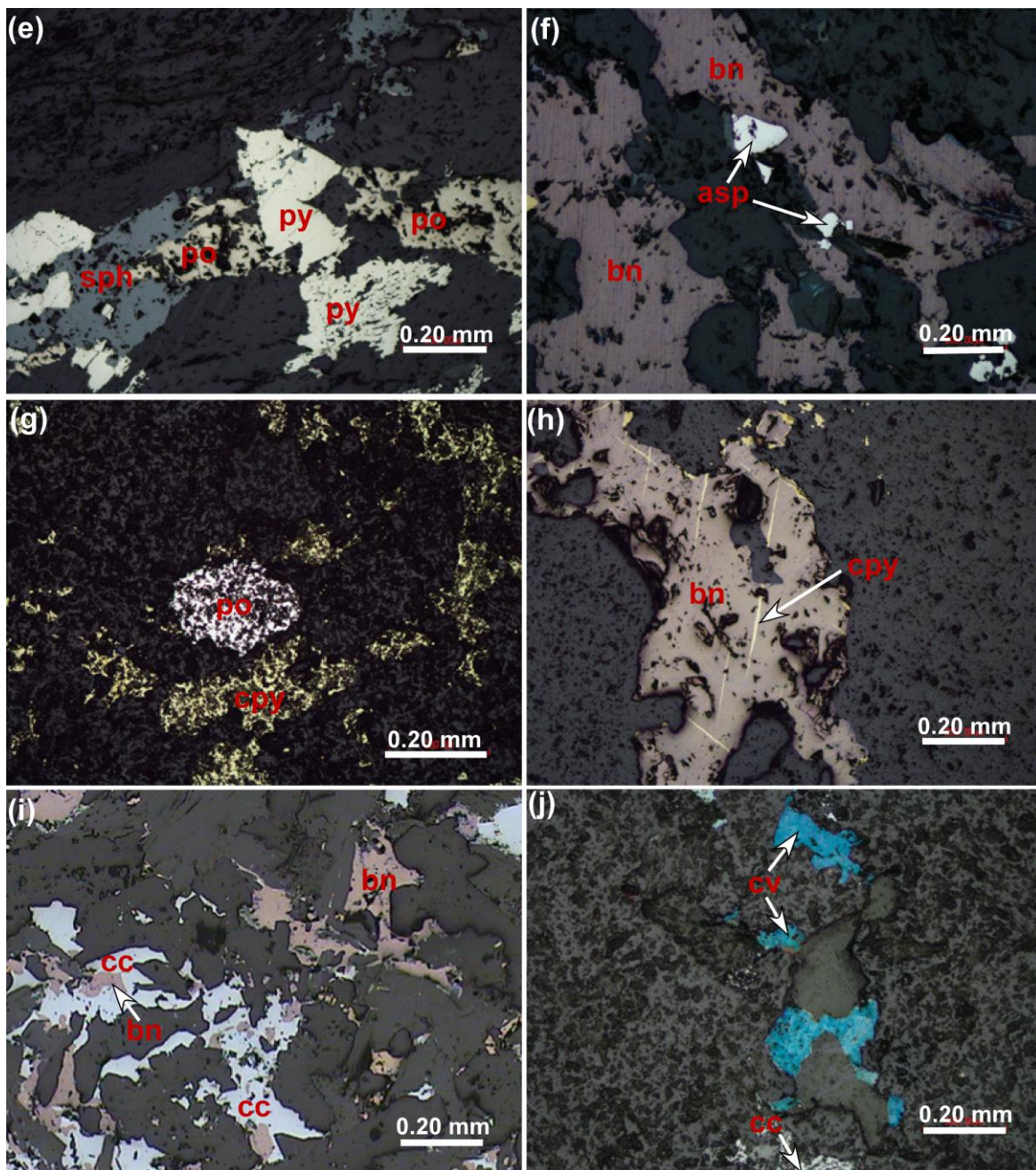


Fig. 5.2 (continued).

5.2. Structure-controlled mineralisation

5.2.1. Vein-hosted mineralisation

In vein-controlled mineralisation style, sulphides are hosted in bedding-parallel, irregular veins/veinlets (Fig. 5.3a, b) and massive veins (Fig. 5.3c) that are concentrated at the base of the D'kar Formation. The bedding-parallel veins are mostly concentrated at the contact between coarse-grained and fine-grained layers. These veins are mm-to-cm thick and are

often folded, fractured, and locally boudinaged, with sulphides concentrated in the neck of the boudins (Fig. 5.3a). The centre of the boudin is occupied mainly by quartz with minor calcite at the margins. The irregular veins range in thickness from hairline size to several centimetres and are unsystematically orientated with respect to the bedding. Locally, the vein sulphide contents are up to 60 vol. % (Fig. 5.4a). Multidirectional sulphide stringers forming stockworks are observed in less competent rocks, i.e. shale. The stringers are thin (0.02-0.2 mm) and up to several millimetres long. The main gangue minerals in these veins are quartz (very coarse grained subhedral to euhedral) and calcite (locally cementing silt sized terrigenous quartz), in variable proportions (calcite only, calcite-quartz, quartz-calcite and quartz only). The minor phases include biotite, muscovite, and chlorite. This style of mineralisation has been documented in similar deposits in the CAC (see **Brems et al., 2009; Torremans et al., 2013, 2014**). In addition to the previously mentioned discrimination criteria (see Chapter 5), veins can be discriminated on the basis of their ore mineralogy: (i) Fe-S veins, (ii) Cu-Fe-S veins, (iii) Cu-Pb-Zn veins, and (iv) iron oxides veins.

The Fe-S veins are characterised by pyrite and marcasite that have a framboidal texture. The thickness of these veins ranges from hairline cracks (Fig. 5.4b) to millimetre- scale. In the Cu-Fe-S veins, the sulphides are dominantly chalcopyrite, bornite, and chalcocite with minor pyrite, pyrrhotite, arsenopyrite sphalerite, galena, cobaltite, and tetrahedrite/tennantite. Pyrite, pyrrhotite, arsenopyrite, cobaltite, sphalerite (Fig. 5.4c), and tetrahedrite/tennantite (Fig. 5.4d) occur as inclusions either in chalcopyrite or bornite. Locally, isolated grains of tetrahedrite/tennantite are observed. The relationship between chalcopyrite, bornite, and chalcocite is not clear as these phases appear to replace one another, form exsolution textures (Fig. 5.4e) and in places chalcocite forms rims around bornite (Fig. 5.4a). In some cases in grey reduced sandstone within the redbed dominated sequence of the Ngwako Pan Formation, sulphides (chalcocite-bornite) are observed in irregular fractures and veinlets and in worm-shaped bodies (Fig. 5.3b). Textural relationship shows that native bismuth (bright white with brownish tarnish) is overgrown by bismuth-bearing sulphides (light yellow to dark brown) possibly emplectite (CuBiS_2) and/or wittichenite (Cu_3BiS_3) that are in turn overgrown by chalcocite that also overgrows bornite (Fig. 5.4f). Rutile (light grey and acicular) also occurs as inclusions in chalcocite and bornite.

The Cu-Pb-Zn veins are characterised by simple ore and gangue mineralogy. The major ore minerals are sphalerite and galena, with minor chalcopyrite. The dominant gangue

mineral is calcite with minor amount of quartz. Sphalerite often overgrows galena and chalcopyrite (Fig. 5.4g) and forms intergrowth with chalcopyrite and bornite (Fig. 5.4h). In addition, magnetite replaces hematite (Fig. 5.4i) and pseudomorphs of pyrite after either biotite or anhydrite/gypsum (Fig. 5.4j) are observed in calcite-anhydrite/gypsum (?) veins hosted by the Ngwako Pan Formation sandstone.

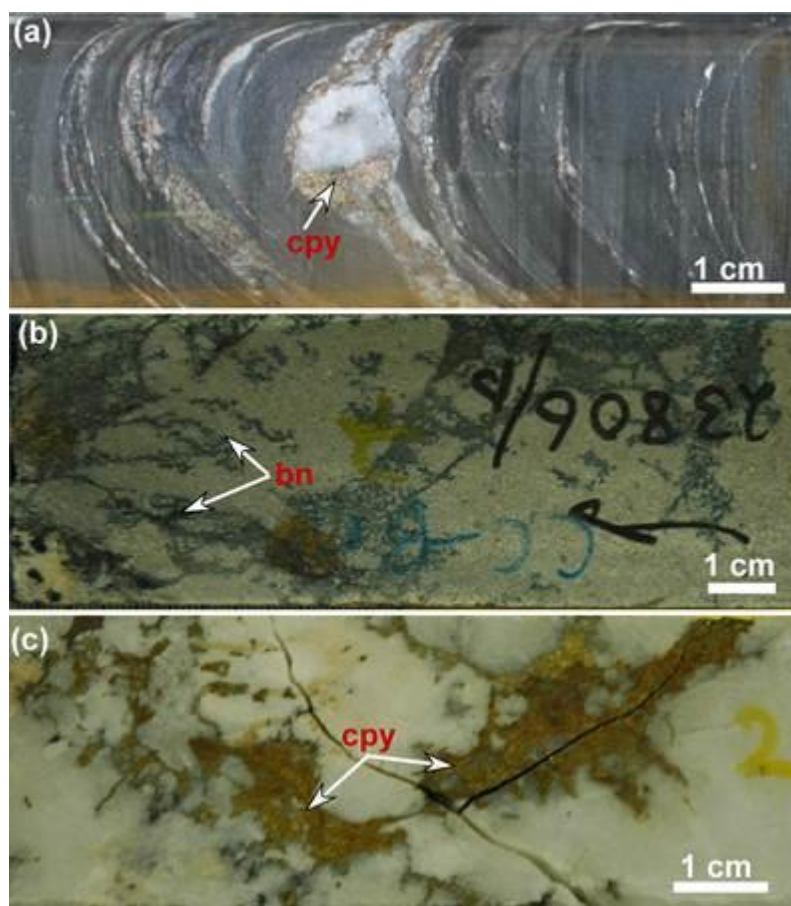


Fig. 5.3: Photographs of drill cores from the Banana Zone showing different vein types hosting copper mineralisation. (a) Chalcopyrite (cpy) concentrated in the neck of a boudin in a boudinage vein (drill core HA 248-D; 205 m). (b) Irregular veinlets and worm-shaped bodies containing bornite in a grey reduced porous sandstone of Ngwako Pan Formation (drill core HA 482-D; 189.10-189.23 m). (c) Chalcopyrite (cpy) in very coarse crystalline quartz-calcite vein (drill core HA 483-D; 116.09-116.29 m)

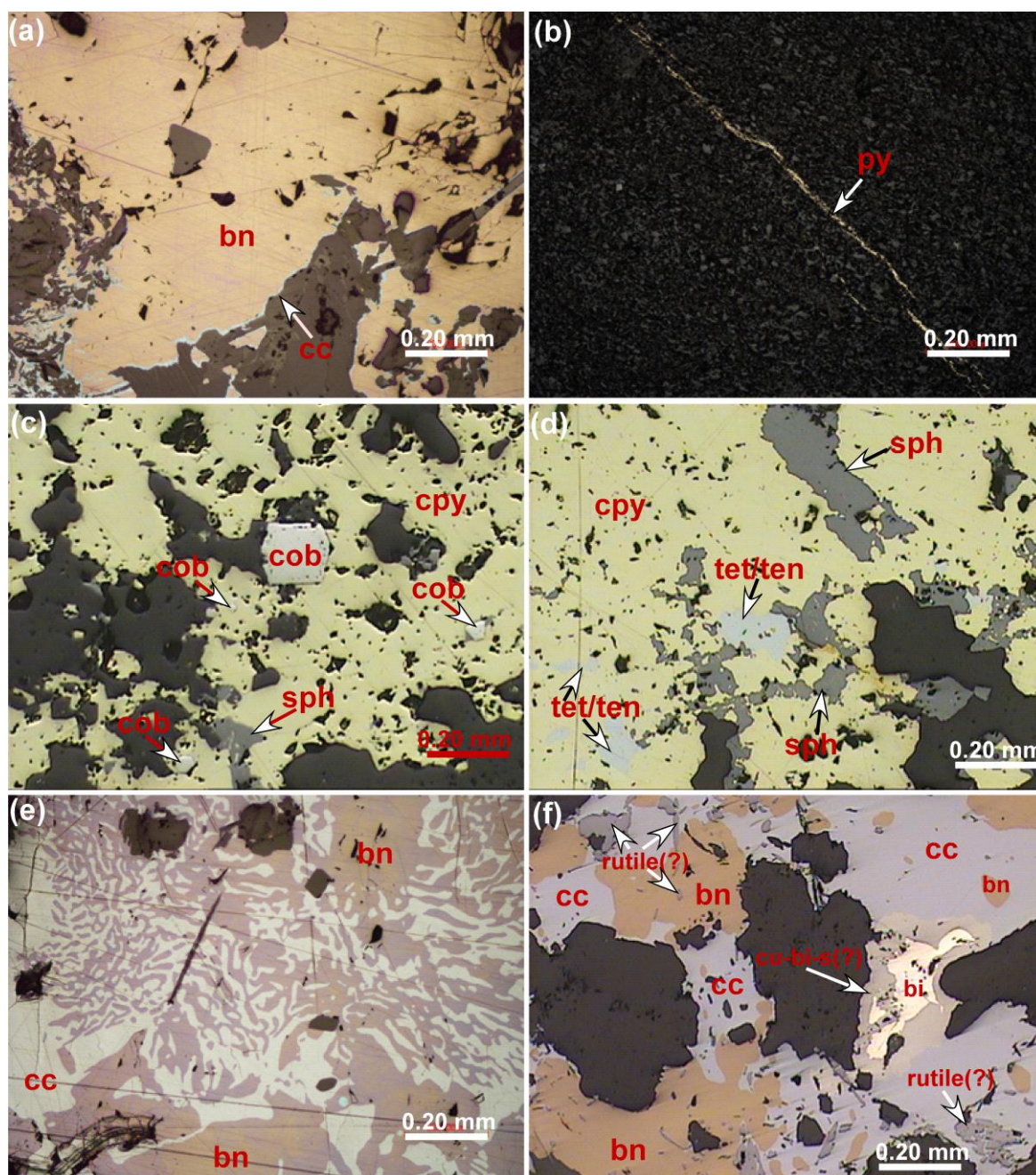


Fig. 5.4: Photomicrographs of polished thin sections (reflected light) showing different vein mineralisation occurrences at the Banana Zone. (a) Irregular vein with a massive xenomorphic bornite (bn) crystal overgrown by chalcocite (cc). (b) Hairline veinlet filled by pyrite (py). (c) Cobaltite (cob) inclusions in porous massive chalcopyrite (cpy). (d) Massive chalcopyrite (cpy) with tennantite/tetrahedrite (ten/tet) and sphalerite (sph) inclusions. (e) Myrmekitic intergrowth of chalcocite (cc) and bornite (bn). (f) Textural relationship between native bismuth (bi), chalcocite (cc), copper bismuth sulphides (?) (cu-bi-s(?)), bornite (bn) and rutile(?). (g) Massive sphalerite (sph) enclosing chalcopyrite (cpy) and galena (gn). (h) Intergrowth texture between sphalerite (sph), chalcopyrite (cpy), and bornite (bn). (i) Replacement of magnetite (mt) by hematite (hm). (j) Pseudomorphic replacement of either biotite or anhydrite crystals by pyrite.

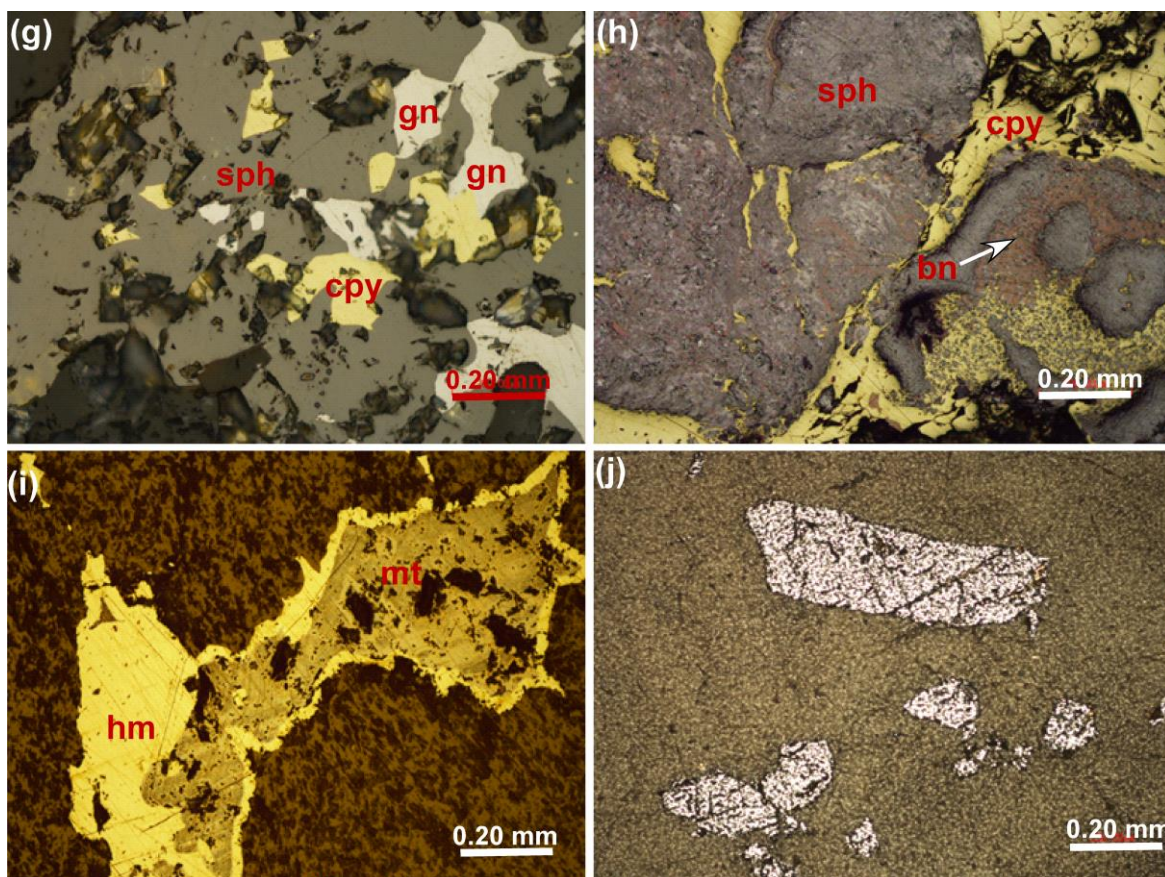


Fig. 5.4 (continued).

5.2.2. Breccia-related mineralisation

In the breccias (Fig. 5.5), sulphides are mainly observed in the matrix/cement encrusting broken fragments of calcite, quartz, and lithics. The matrix sulphide minerals include chalcopyrite and bornite.

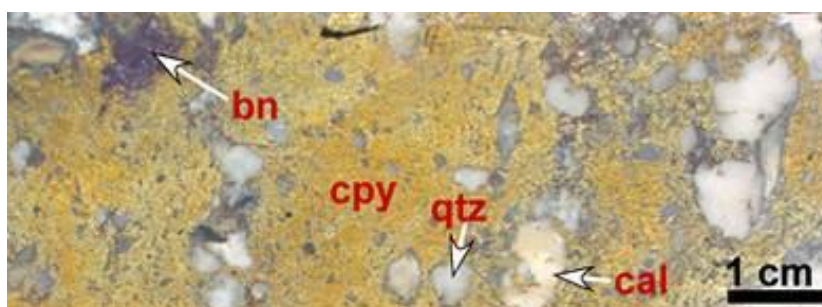


Fig. 5.5: Photograph of a drill core from the Banana Zone showing breccia-related mineralisation; sulphide mineralisation (bn and cpy) cements quartz and calcite clasts (drill core HA 10-D; 160.03-160.14 m).

5.2.3. Cleavage-controlled mineralisation

Cleavage-controlled mineralisation style represents an additional structurally-controlled mineralisation style observed in the Banana Zone. This style of mineralisation is observed in cleavage of rocks (Fig. 5.6a, b), as well as in cleavage of minerals (Fig. 5.7a, b). Cleavage-controlled mineralisation style is essentially observed in shear zones within the shale, where two sets of cleavage planes filled by bornite are observed. The primary cleavage is concordant to bedding whereas the secondary cleavage crosscut both the bedding and the primary cleavage. Also, in deformed marlstone that had developed crenulation cleavage bornite mineralisation is observed to be transposed along the cleavage traces (Fig. 5.6a, b). At mineral-scale calcite is preferentially replaced by pyrite along cleavage planes. The replacements have occurred preferentially along the mineral cleavage planes (Fig. 5.7a, b) which have likely channelised mineralising fluids.

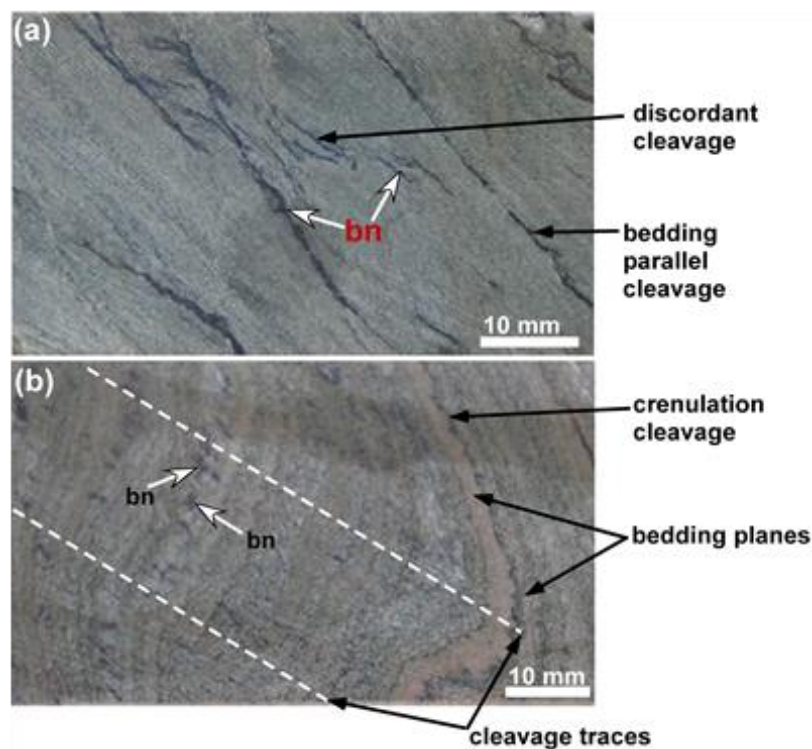


Fig. 5.6: Photographs of drill cores showing cleavage-hosted of mineralisation from the Banana Zone. (a) Bornite (bn) mineralisation transposed along two sets of cleavages (drill core HA 487-D; 127.49-127.58 m). (b) Marlstone unit presenting crenulation cleavage parallel to the stratification and hosting bornite (bn) mineralisation transposed along cleavage traces (drill core HA 520-D; 164.76-164.93 m).

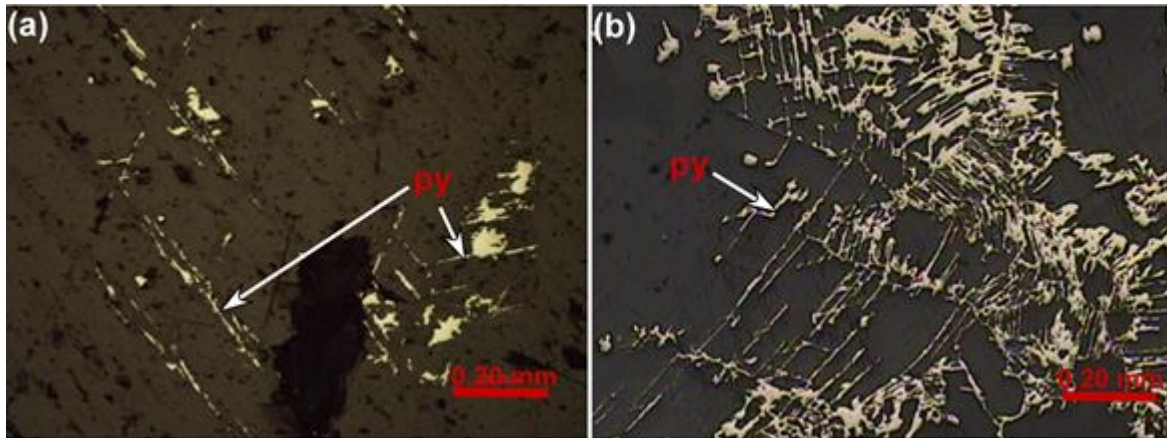


Fig. 5.7: Photomicrographs of polished thin sections (reflected light) showing Pyrite (py) replacing calcite preferentially along cleavage planes

From the above results, it can be concluded that the complex ore-forming processes involve multiple stages that span from diagenesis through metamorphism and deformation to post-tectonic.

CHAPTER 6

WALL ROCKS ALTERATION

The mineralisation at the Banana Zone is concentrated in very fine-grained sedimentary rocks, i.e., shale and siltstone, hence difficulties in observing the alteration at the field scale. Furthermore, these rocks have undergone low grade regional metamorphism making it difficult to discriminate between hydrothermal alteration assemblages and metamorphic mineral assemblages. However, metamorphic minerals are distinct in their large grain size and higher modal abundance. On the other hand, the alteration minerals lack any preferred orientation, indicating they are of hydrothermal and not of regional metamorphism origin. Two modes of occurrences of hydrothermal alteration were identified and these include the pervasive and the vein-controlled modes as summarised in Table 6.1. In addition to hydrothermal alteration, wall rocks underwent weathering and related supergene enrichment processes. In this context, hydrothermalism refers to any process involving hot fluids (<100 °C to 800 °C, **Ridley, 2015**). Therefore, diagenetic and metamorphic fluids are considered as hydrothermal.

6.1. Pervasive alteration

This form of alteration, which entirely overprints the country rocks, is recognised by a colour change that includes bleaching and darkening. The alteration type observed differs depending mostly on the lithology. The following alteration types were differentiated on the basis of the alteration mineral assemblages.

6.1.1. Sodic-calcic alteration

This alteration type with characteristic white to pink appearance is widespread and particularly well developed in the medium-grained sandstone of the upper section of the D'Kar Formation. Sodic-calcic alteration is characterised by mineral assemblage including albite, calcite, and chlorite. X-ray diffractometry (XRD) analysis (see Appendix A3) shows that the most abundant alteration mineral phases are quartz (33 wt.%), albite (25

wt.%), calcite (20 wt.%) and chlorite (16 wt.%), while the content of muscovite is minor (~ 6 wt.%). The predominance of albite and calcite gives the rock a bleached appearance. It is important to note that not all the quartz observed here is of hydrothermal origin as demonstrated by presence of rounded quartz that may be of detrital origin. Though sphalerite and galena are found to be associated with the sodic-calcic alteration, field observations show that the mineralisation was most likely brought by carbonate veining that overprinted the sodic-calcic alteration (Fig. 6.1). The sphalerite encloses tiny inclusions of chalcopryite, a phenomenon known as chalcopryite disease (**Ramdohr, 1969; Barton, 1978; Barton and Bethke, 1987**). The sodic-calcic alteration type that can be regarded as the inverse of potassic alteration (**Carten, 1986**) occurs as a result of the metasomatic introduction of sodium and calcium carbonate solutions.

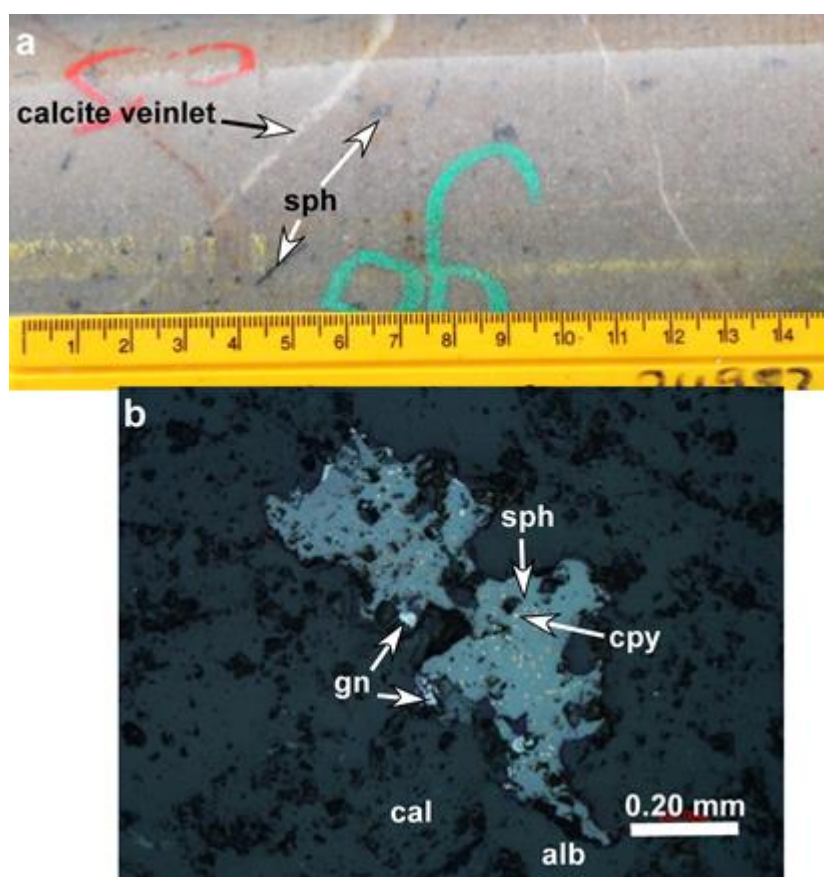


Fig. 6.1. (a), Photograph showing a pervasive sodic-calcic alteration dominated by albite and calcite assemblage in the sandstone unit from the D'kar Formation (drill core HA 487-D; 72.04-72.17m). Note a carbonate veinlet overprinting the sodic-calcic alteration. The tiny black spots are sphalerite and galena that has no association with the sodic-calcic alteration, but rather associated with the carbonate overprint. (b), Plane polarised reflected light photomicrograph showing sphalerite (sph) and galena (gn) associated with calcite (cal) and albite (alb) assemblage. Sphalerite (sph) contains tiny inclusion of chalcopryite (cpy).

6.1.2. Phyllic alteration

This alteration type is observed in fine-grained rocks, i.e., siltstone and mudstone. The alteration is pervasive and totally overprints the rock, which is bleached (Fig. 6.2a). The mineral assemblages characterising this alteration style include muscovite, quartz, chlorite, albite, and locally traces of sulphides. XRD analyses (see Appendixes A2 and A3) show that muscovite (2M₁ variant) (35-47 wt.%) is the most abundant alteration mineral phase, followed by quartz (26-30 wt.%), Mg-rich chlorite (16-34 wt.%), and albite (2-14 wt.%). The muscovite is a result of the alteration of feldspars, a phenomenon widely observed petrographically (Fig. 6.2b). Muscovite is characterised by very fine-grained texture and was therefore referred to as sericite in petrographic investigations. Quartz encloses both chlorite and sulphides, which in turn enclose muscovite. Therefore, chlorite and sulphides are cogenetic, but with chlorite precipitating relatively later than the sulphides as indicated by the inclusions of sulphides in chlorite but no inclusions of chlorite in sulphides. Sulphides associated with this alteration type include pyrite, pyrrhotite, arsenopyrite and minor chalcopyrite. The pyrite is characterised by fine-grained anhedral framboidal grains and coarse-grained, diagenetic euhedral crystals that consist exclusively of framboidal aggregates (see Figs. 5.1b and 5.2c, d in Chapter 5). Locally, sericitised layers are crosscut and displaced by calcite-quartz veinlet (Fig. 6.2c).

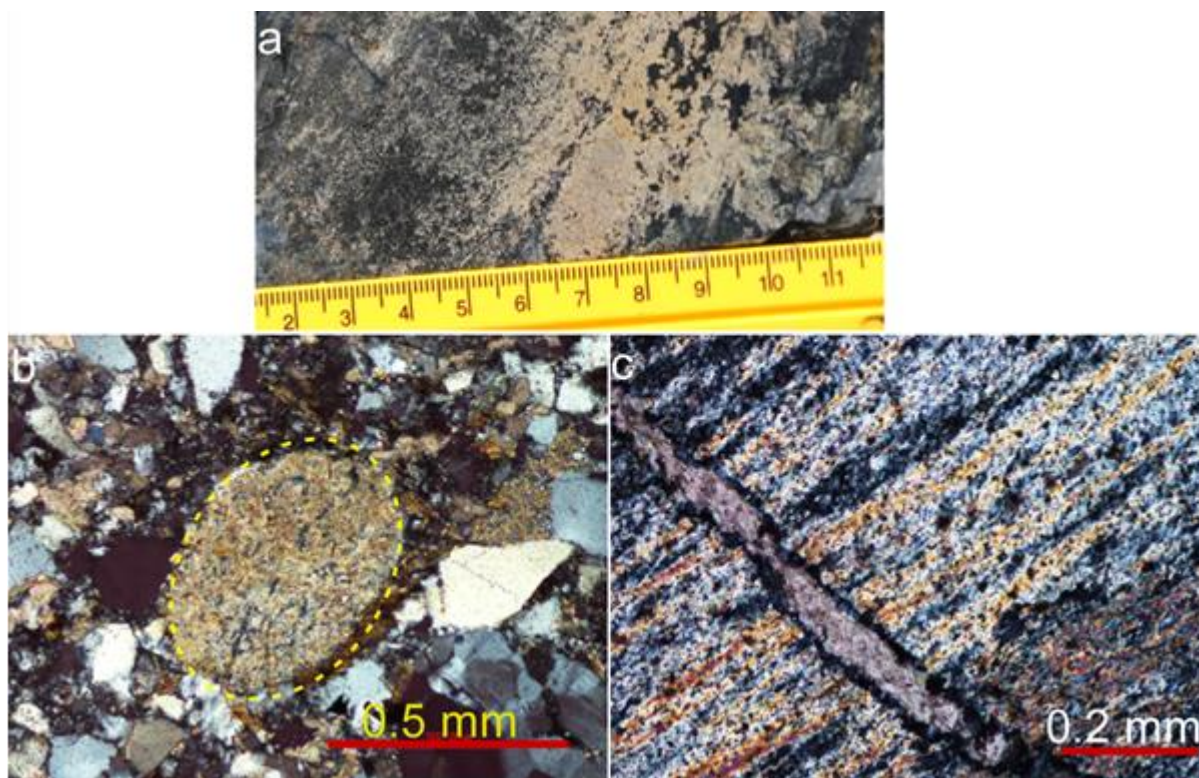


Fig. 6.2. (a), Photograph showing pervasive phyllic alteration in a siltstone unit from the D'kar Formation (drill core HA 244-D; 213.89-213.99 m). (b) Cross polarised transmitted light photomicrograph showing a complete replacement of microcline (yellow circle) by sericite. (c) Cross polarised transmitted light photomicrograph showing the relationship between the sericite-quartz layers cut and displaced by a later calcite-quartz vein.

5.1.3. Silicification

Silicification is recognised by a noticeable increase in the hardness and the bleaching of the rocks. Furthermore, the rocks are characterised by an increase in the proportion of quartz or crypto-crystalline silica and relatively high $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio of *ca* 7 (see Chapter 4). Silicification is manifested as a pervasive quartz replacement. This alteration is often accompanied by the dissolution of the rocks creating voids in the rock (Fig. 6.3). Locally, the original texture of the rock is completely destroyed (Fig. 6.7). Malachite and chalcocite are the associated ore minerals in the Ngwako Pan Formation sandstone, whereas pyrite, chalcopyrite, and digenite accompany this alteration in shale/mudstone of the D'kar Formation (see Fig. 6.3). Of note is that malachite, which is of supergene origin is unlikely genetically associated with silicification and may be the result of overprinting process.

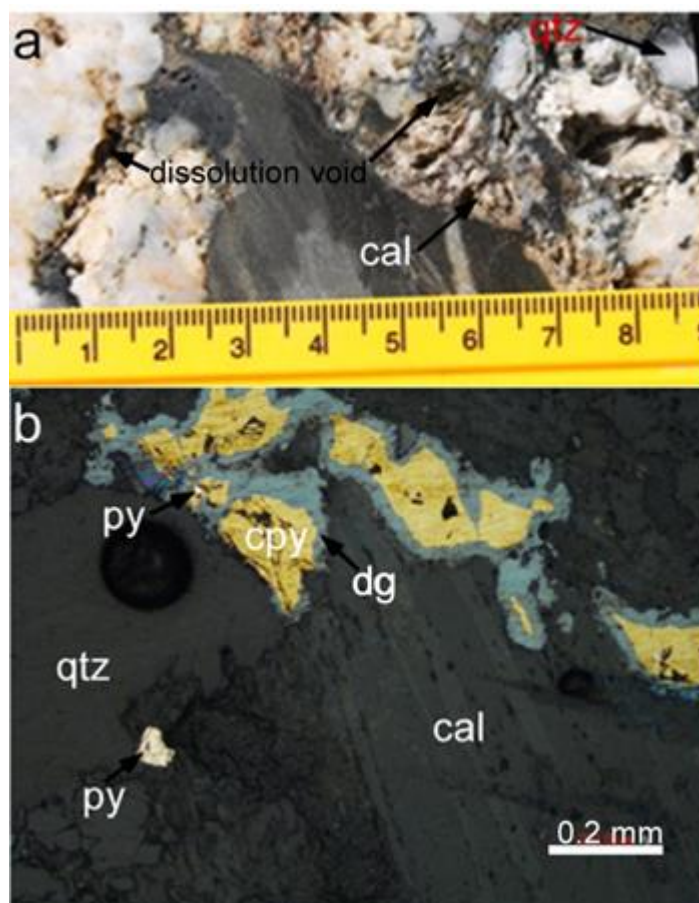


Fig. 6.3. (a) Photograph of a mudstone that has undergone silicification. Dissolution is observed to accompany the silicification process leaving behind voids (drill core HA 90-D; 104.33-104.40 m). (b) Reflected light photomicrograph showing pyrite (py) overgrown by chalcopyrite (cpy) that is rimmed by digenite (dg) in a silicified rock. The mineralisation is associated with quartz (qtz) and calcite (cal) phases.

6.1.4. Propylitic alteration

This alteration type is recognised by the green to dark greenish colour of the rocks, due to the dominance of chlorite (Fig. 6.4). This alteration type, which is essentially indistinguishable from the assemblages that form during regional greenschist metamorphism is characterised by chlorite-albite-calcite-quartz-muscovite $\pm$ anhydrite mineral assemblage. It occurs as either pervasive replacement of mafic minerals or from the introduction of magnesium and iron (**Evans, 1993**), thus resulting in the dominance of chlorite. XRD analysis (see Appendixes A2 and A3) shows that Mg-rich chlorite (clinochlore) (35 wt.%) is the most abundant alteration mineral, followed by albite and calcite (21 wt.% each) quartz and muscovite (2M₁) (11 wt.% each) and traces amount of anhydrite. Epidote is absent or scarce. The ore minerals associated with this alteration type

include chalcopyrite, bornite, and chalcocite. There is a clear association between sulphide mineralisation and chlorite, as demonstrated by the inclusion of one in the other.

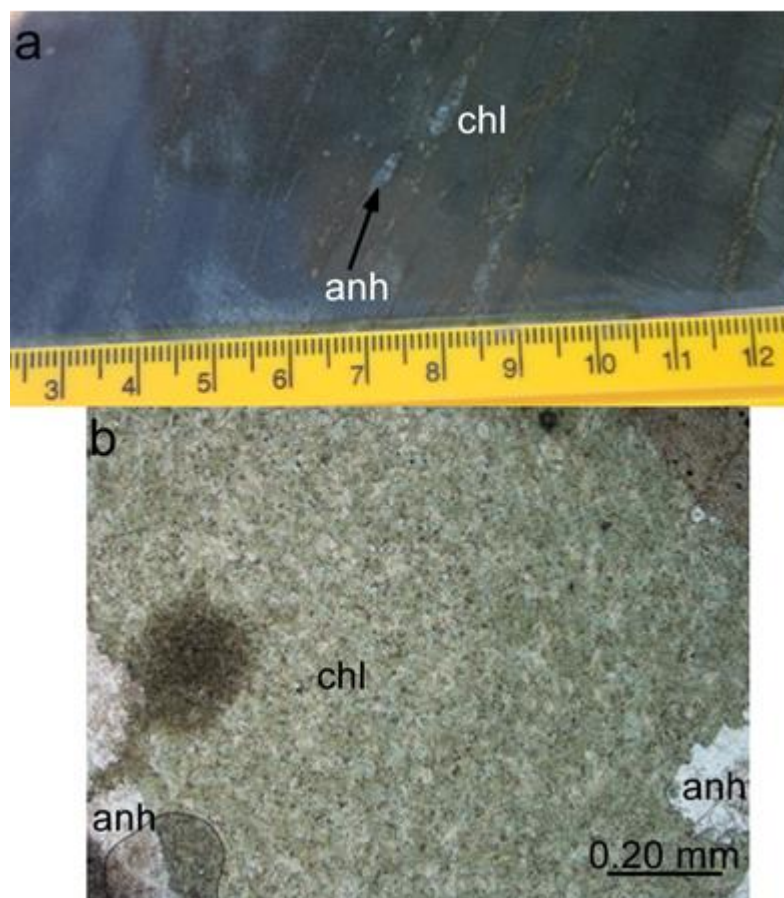


Fig. 6.4: (a) Photograph of a propylitic-altered mudstone rock (drill core HA 524-D; 331.34-331.46 m) showing pervasive propylitisation of the mudstone. (b) Photomicrograph of pervasively chloritized mudstone, transmitted plane polarized light. anh = anhydrite; chl= chlorite.

6.1.5. Saussuritisation

Another type of alteration associated with coarser grained clastic lithologies, i.e., the sandstone of the Ngwako Pan Formation is saussuritization. This alteration is characterised by saussurite mineral assemblage of chlorite-sericite-calcite that indiscriminately replace both the clasts and the matrix. The clasts, specifically the calcium-rich plagioclase are partially or completely altered to saussurite (Fig. 6.5). The ore minerals observed in the saussuritised clasts are chalcocite and bornite. The relationship between the chalcocite and the bornite suggests cogenetic crystallisation as indicated by inclusions of one in the other.

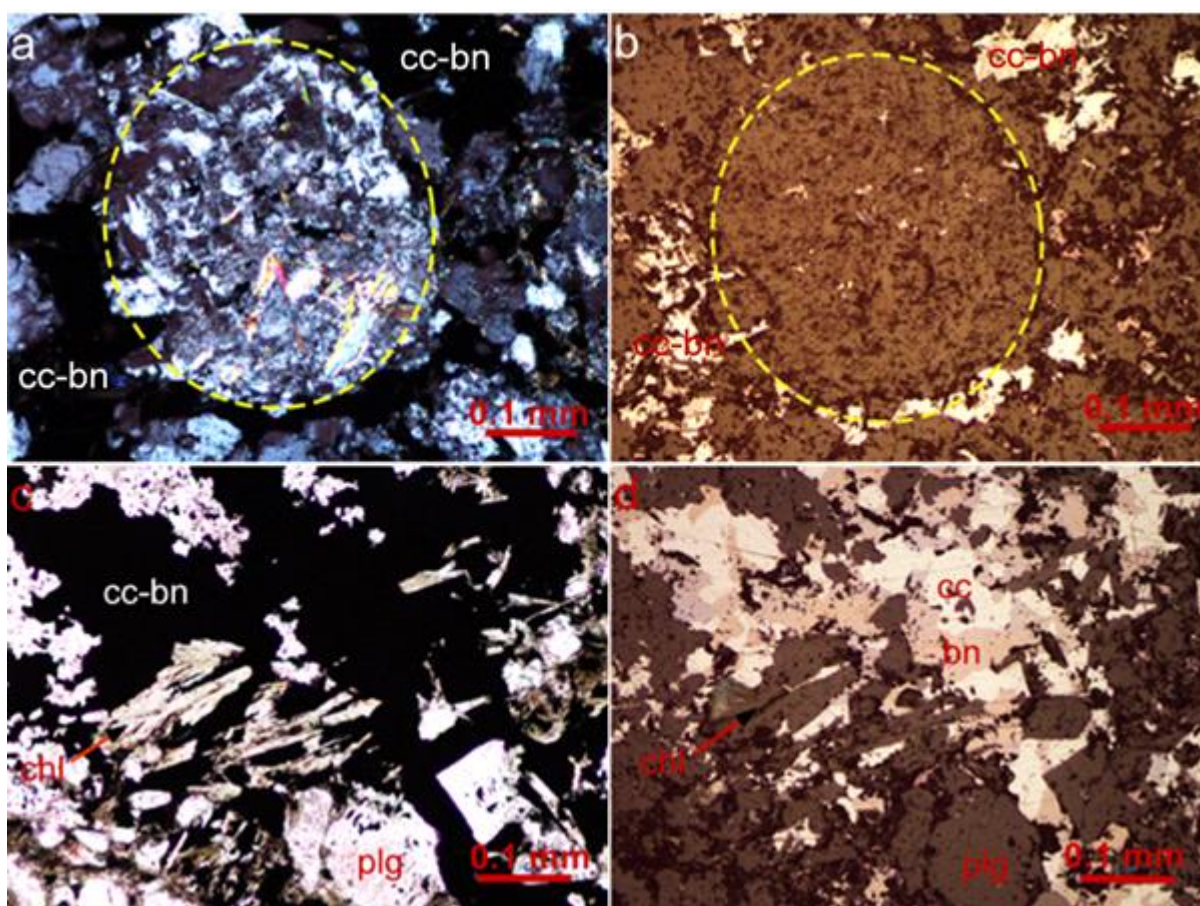


Fig. 6.5. Photomicrographs illustrating saussurite assemblage in Ngwako Pan Formation sandstone (a) and (b) Replacement of plagioclase (plg) clast (yellow circle) by zoisite, chlorite, calcite, and tiny specks of chalcocite-bornite mineralisation (c), (d) Replacement of early formed cement by chlorite (chl) and chalcocite (cc)-bornite (bn) mineralisation (a, transmitted light, (XPL); b, reflected light). (c, transmitted light, (PPL); d, reflected light).

6.1.6. Carbonate $\pm$ quartz alteration

This alteration type is characterised by the whitened (bleached) zones. This alteration totally overprints the rock (Fig. 6.6) and is widespread in most of the lithologies that host ore minerals. This alteration type is characterised broadly by a calcite-quartz-sulphides $\pm$ muscovite $\pm$ chlorite $\pm$ biotite $\pm$ albite mineral assemblage. The most abundant alteration mineral as indicated by XRD data (see Appendix A3) is calcite (40-92 wt.%), followed by quartz (4-33 wt.%), chlorite (<1-14 wt.%), muscovite (3-9 wt.%), and albite (0-2 wt.%). It has to be noted that not all the quartz observed here is of hydrothermal origin as demonstrated by presence of rounded quartz that may be of detrital origin. Petrographic study shows that the calcite crystals are coarse-grained and exhibit twinning (Fig. 6.6). Sulphides, quartz, muscovite, and biotite occur as inclusions in calcite. Calcite is enclosed

within the chlorite, thus indicating that calcite is earlier than chlorite. Sulphides associated with this alteration type are usually occurring as disseminations and streaks or lenses and include mostly chalcopyrite, bornite, and chalcocite with minor occurrence of sphalerite and galena. Most of the Cu and Cu-Fe sulphides mineralisation in the Banana Zone occurs predominantly within zones of carbonate-quartz alteration.

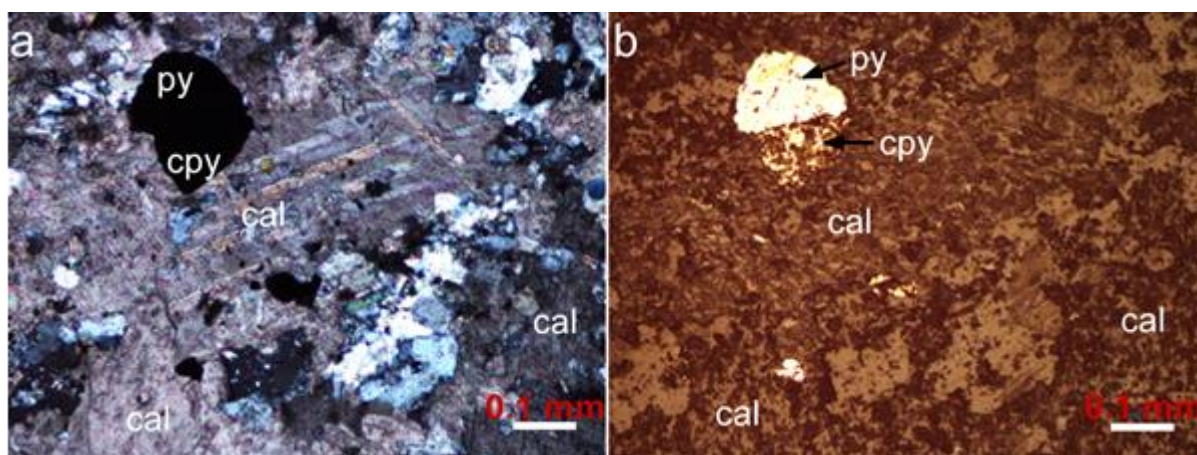


Fig. 6.6. Photomicrographs showing pervasive carbonate alteration overprinting the earlier formed hydrothermal minerals. (a), (b) Calcite (cal) enclosing the detrital grains, chalcopyrite (cpy) and pyrite (py). (a), cross polarised transmitted light (XPL); (b), XPL reflected light.

6.1.7. Supergene enrichment

The sandstone of the D'Kar Formation has undergone weathering and supergene enrichment. This process is widespread in the Chalcocite subzone of the Banana Zone (see the location map, Fig. 1.1), where extensive transformation of hypogene chalcocite to secondary chalcocite is a common phenomenon. This secondary chalcocite differs from the primary one by the fact that it is characterised by the absence of intergrowths with hypogene bornite or chalcopyrite and is often associated with secondary porosity (cf. **Torremans et al., 2013**). In zones where chalcocite is abundant, the adjacent host rocks are characterised by the reddish colour due to the presence of iron oxyhydroxides. Near the surface, oxidation of sulphides by circulating meteoric fluids resulted in the precipitation of malachite and hematite in dissolution cavities (Fig. 6.7). Covellisation of hypogene copper and copper-iron sulphides is also observed.



Fig. 6.7. Photograph showing supergene mineralisation in a silicified rock. cc; chalcocite, qtz; quartz, hm; hematite, mal; malachite (drill core HA 487-D; 144.13-144.22 m). Note; the primary texture of the rock is completely destroyed; malachite is not genetically related to the silicification.

6.2. Vein-controlled form of alteration

All lithological units are crosscut by veins and veinlets. These veins and veinlets can be barren or mineralised; they can be with or without alteration halos or selvages. Veins can be discriminated in several ways that include (i) alteration mineral assemblage, (ii) ore mineralogy, (iii) the mode of occurrence, and (iv) state of deformation. The following is a description of some of the most common types based on their mineral assemblage.

6.2.1. Carbonate-quartz-sulphides $\pm$ chlorite $\pm$ biotite $\pm$ muscovite veins

These veins, which are either deformed or undeformed, have a thickness of 0.2 to 3 cm, and contain calcite and quartz. The amount of calcite and quartz in these veins varies considerably. In some veins calcite is the dominate phase, whereas in others the main phase is quartz. The deformed veins occur in two forms that include boudinaged veins and brecciated veins. Calcite is the main gangue mineral observed in these veins. A syntaxial texture is displayed within the boudinaged veins, with quartz commonly occurring at the centre of the boudin, while calcite and a small quantity of chlorite occur at the margin. The calcite in these veins is deformed as demonstrated by the presence of bent laminae. Sulphides that seem not to be deformed are concentrated in the neck of the boudins (Fig. 6.8a-c). In the brecciated veins, sulphides occur as cement, in which calcite, quartz and feldspar clasts set. The feldspar clasts have undergone saussuritisation and enclose

sulphide grains (Fig. 6.9a, b). Inclusions of quartz were observed in calcite crystals. The sulphide species associated with calcite-quartz $\pm$ chlorite veins in both brecciated and boudinage veins are comprised of chalcopryite and bornite. The relationship between these sulphides suggests cogenetic crystallisation as indicated by bornite inclusions in chalcopryite and vice versa.

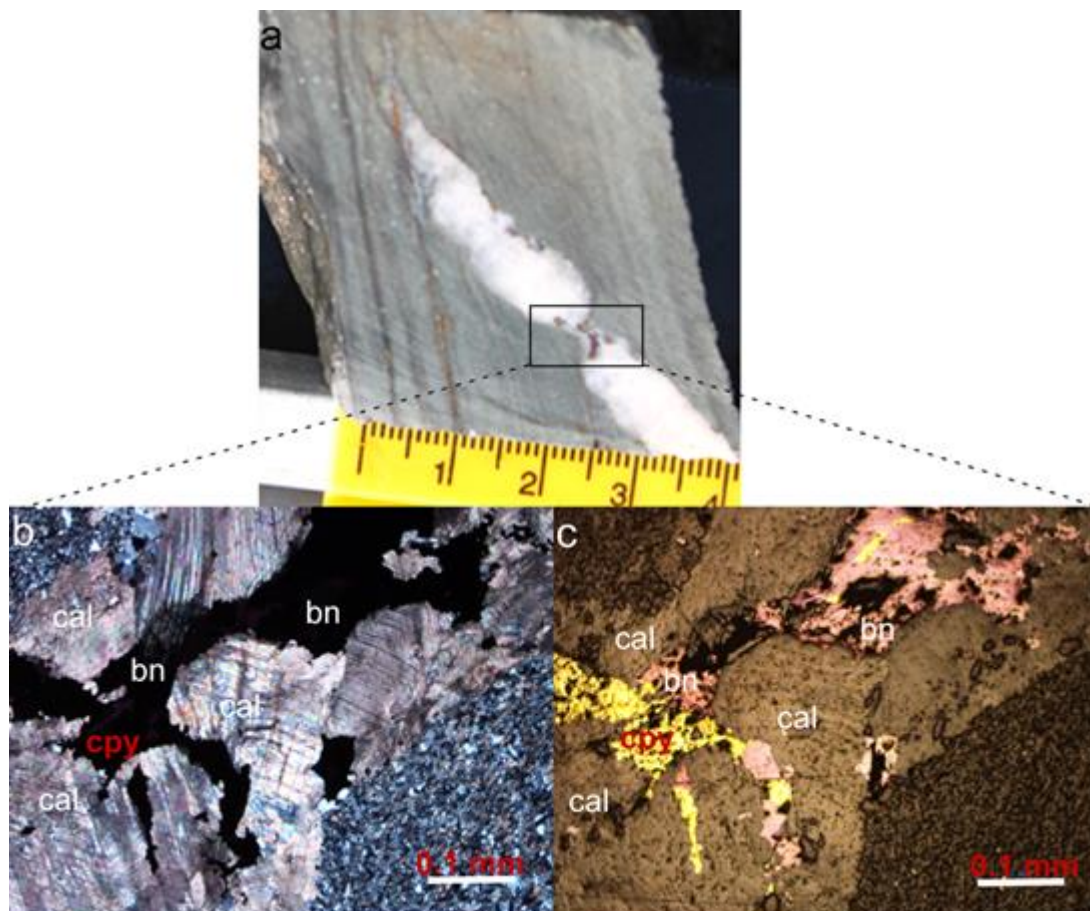


Fig. 6.8. (a), Photograph showing boudinage calcite (white area) veinlet that is oblique to the laminations (drill core HA 244-D; 203.38-203.42 m). Chalcopryite and bornite (enclosed in a rectangle) mineralisation concentrated in the neck of the boudin. (b), XPL transmitted light photomicrograph of the selected area. (c), Reflected light photomicrograph of (b). bn= bornite; cpy= chalcopryite.

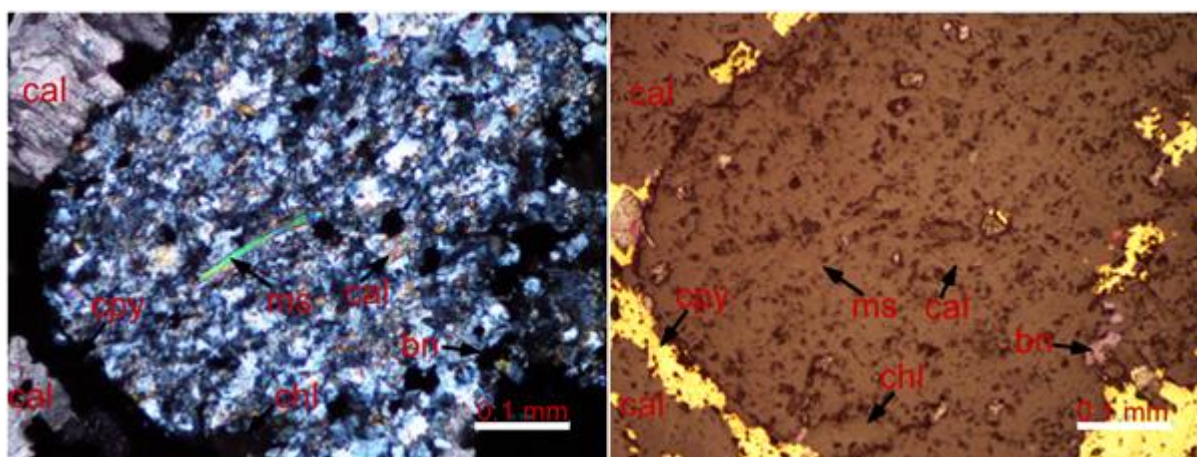


Fig. 6.9. (a), XPL Transmitted light photomicrograph of plagioclase clast in a brecciated calcite-quartz vein being altered to saussurite that include muscovite (ms), calcite (cal), and chlorite (chl) assemblage. Chalcopyrite (cpy) and bornite (bn) mineralisation occurs as inclusions in the plagioclase as well as cement in brecciated calcite-quartz vein. (b), Reflected light microphotograph of (a).

In some cases, these veins are undeformed and parallel to the stratification. In these vein types, quartz which is the main gangue mineral is localised along the margins of the vein whereas the centre of the vein is occupied by calcite and sulphides. Locally, small quantities of chlorite and/or muscovite were observed. Chlorite and biotite are engulfed by quartz, which in turn forms rims around sulphides, and inclusions of sulphides are observed in calcite and biotite. The accompanying sulphides include pyrite, pyrrhotite, galena, sphalerite, and chalcopyrite. Pyrrhotite forms incomplete rims around pyrite (Fig. 6.10). Pyrite, pyrrhotite, galena, and sphalerite are enclosed within the chalcopyrite. The relationship between quartz and chalcopyrite suggests contemporaneous crystallisation as indicated by the inclusions of one in the other. This vein type is characterised by a narrow (~ 0.4 mm) alteration halo composed of sericite, quartz, and chlorite.

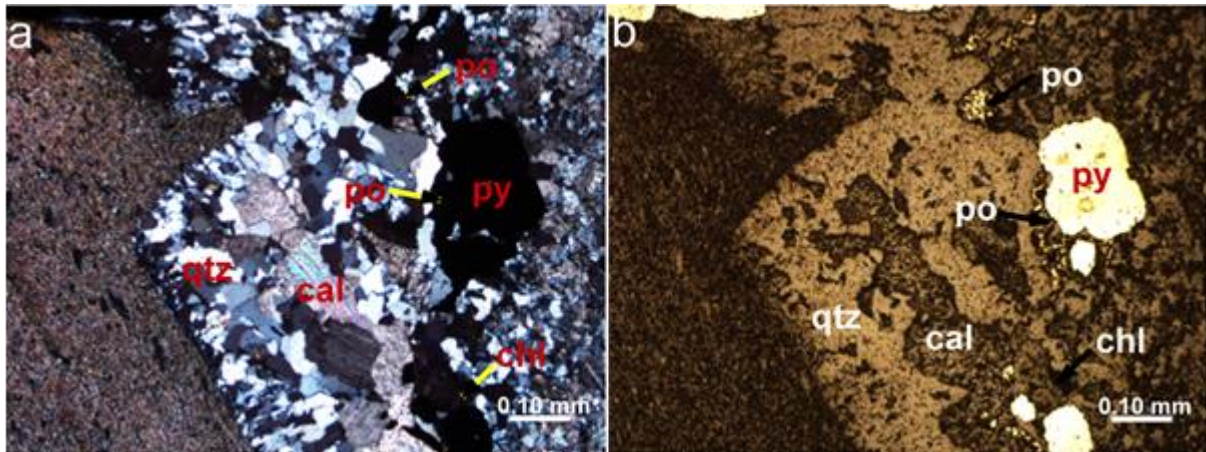


Fig. 6.10. Photomicrographs of quartz (qtz)-calcite (cal) $\pm$ chlorite (chl) vein associated with pyrite (py) and pyrrhotite (po) mineralisation. Pyrrhotite (po) often rims pyrite (py). (a), transmitted light (XPL); (b), reflected light.

6.2.2. Carbonate-quartz-chlorite-hematite veins

These veins are concordant to the stratification and crosscut the propylitic-altered D'Kar Formation shale unit (Fig. 6.11). The thickness of these veins range from 0.1 to 6 cm (average 2 cm). The original morphology of these veins has been modified by extensional strain resulting in boudinage features. Calcite, which is the dominant gangue phase, is observed in the centre of the veins, whereas at the margins, quartz and hematite are observed. Quartz and chlorite are enclosed in hematite demonstrating that these two phases are earlier than hematite. Calcite and hematite are cogenetic as demonstrated by straight grain boundaries between the two phases, whereas the relationship between quartz and chlorite is unclear.

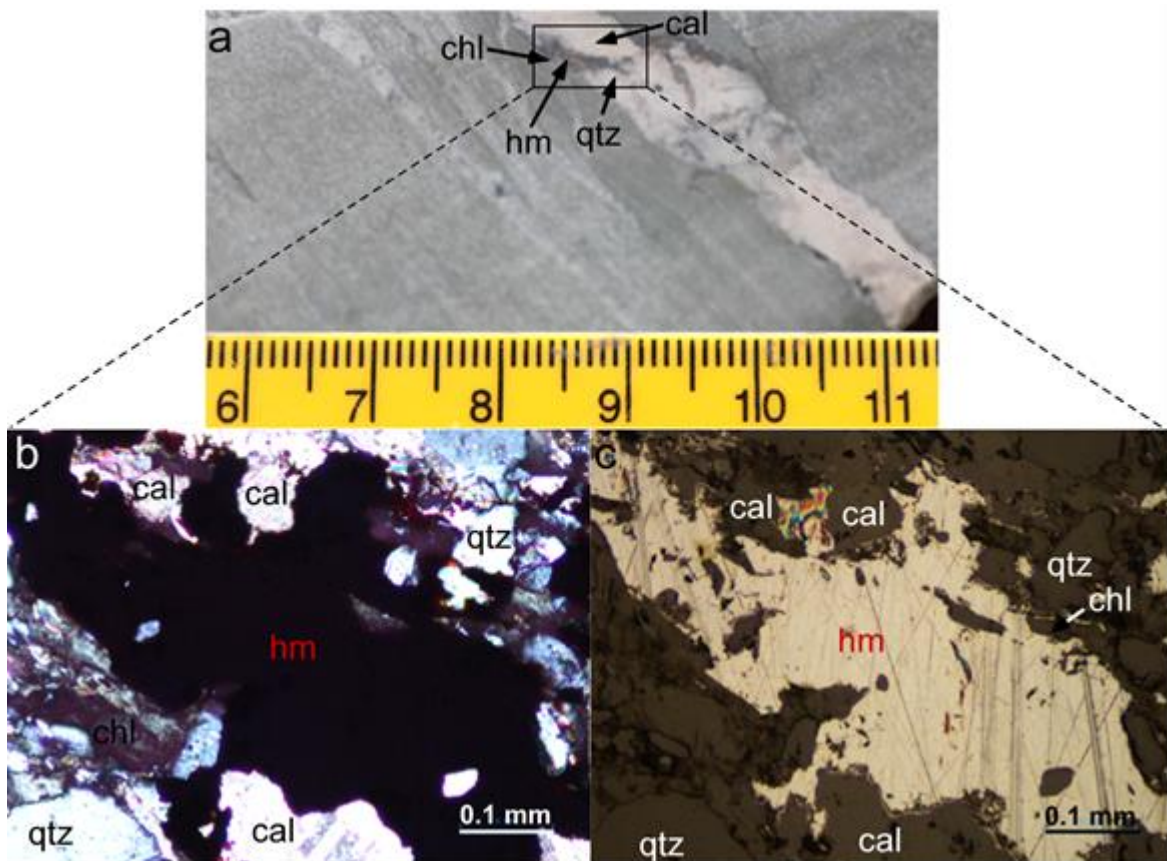


Fig. 6.11. (a) Photograph of deformed, bedding-parallel calcite (cal)-quartz (qtz)-hematite (hm)±chlorite (chl) vein (drill core HA 520-D; 166.21-166.32m). (b), XPL transmitted light photomicrograph of the selected area. (c), Reflected light photomicrograph of (b).

In the oxidised sandstone of the Ngwako Pan Formation, this vein type is spatially associated with iron oxide mineralisation that includes hematite that overgrows magnetite (Fig. 6.12a). In this instance, the vein is undeformed and discordant to the stratification. Early hematite is observed along the vein margins. The hematite occurs as blades and locally fills in the open space between the gangue minerals. Textural associations demonstrate that the calcite and oxide phases are non-cogenetic as shown by irregular and protruding boundaries (Fig. 6.12b). This relationship shows that calcite precipitated earlier than oxides. However, the quartz and the oxides have straight crystal boundaries (Fig. 6.12c) suggesting that the two phases are in a perfect textural equilibrium.

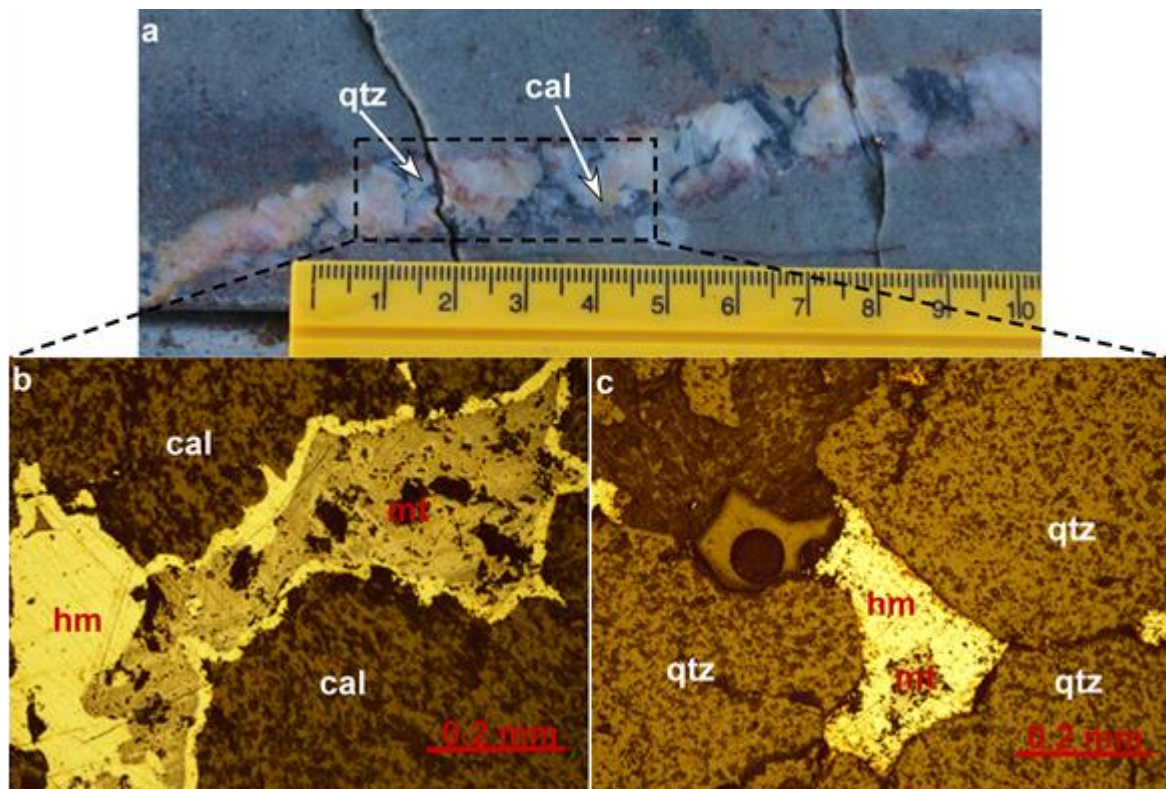


Fig. 6.12.(a), Photograph of the Ngwako Pan Formation sandstone crosscut by quartz-calcite-hematite vein (drill core HA 38-D; 155.30-155.36 m). (b, c), Reflected light photomicrographs showing textural relations between the gangue minerals and hematite (hm) that replaces magnetite (mt) (b), irregular and protruding grain boundary between the calcite (cal) and the hematite (hm), suggesting textural disequilibrium. (c), straight boundary between the quartz (qtz) and the hematite (hm) that infers textural equilibrium between quartz and hematite.

6.2.3. Quartz-sulphides $\pm$ muscovite veins

These veins, which are parallel to the bedding and boudinaged (Fig. 6.13), have a thickness of ~ 0.1 to 2 cm. In these veins, quartz is the main phase, with accessory phases including muscovite (sericite) and sulphides (bornite and chalcopyrite). Bornite that is being replaced by chalcopyrite is enclosed by quartz. Both the gangue and sulphides have undergone brittle deformation.

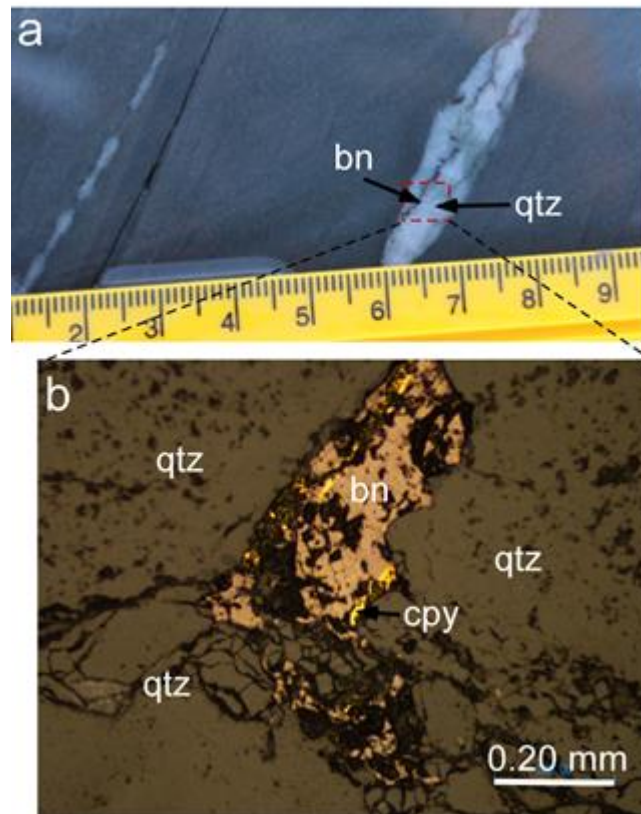


Fig. 6.13. (a), Photograph of a deformed bedding parallel quartz-sulphide vein crosscutting the shale of the D'Kar Formation (drill core HA 38-D; 150.93-151.00 m). (b), Reflected light photomicrograph of selected area in (a) showing the effect of deformation (fracturing) on both the gangue and sulphide minerals. bn = bornite; cpy = chalcopyrite; qtz = quartz.

6.2.4. Calcite-hematite $\pm$ siderite veinlets

The Ngwako Pan Formation sandstone unit is locally crosscut by a narrow (~ 0.5 cm) hematite-bearing calcite veinlet (Fig. 6.14). Hematite occurs in the form of blades that typify specular hematite, which is clearly hydrothermal origin.

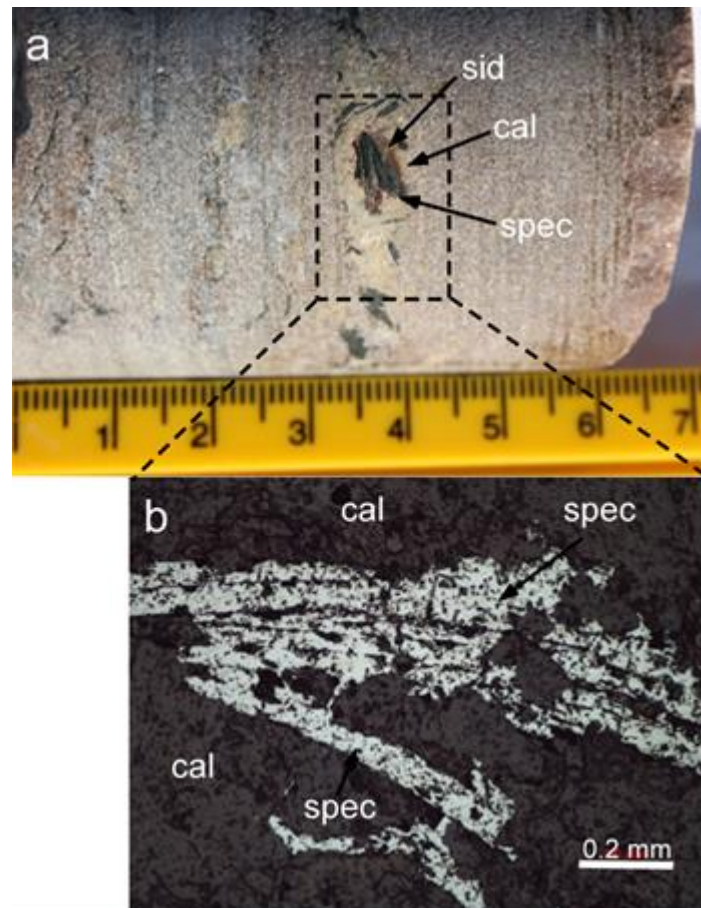


Fig. 6.14 (a) Photograph of specularite (spec) in a calcite (cal)-siderite (sid) veinlet crosscutting the Ngwako Pan Formation sandstone unit (drill core HA 17-D; 245.74-245.80 m). (b) Reflected light photomicrograph showing specularite (spec) in calcite (cal) veinlet.

6.2.5. Quartz-plagioclase-sulphides $\pm$ biotite $\pm$ chlorite veins

These veins, which are undeformed, are characterised by quartz-plagioclase-sulphides $\pm$ biotite $\pm$ chlorite mineral assemblage and measure approximately 1 m in thickness. Quartz is the most abundant alteration mineral and it fills up almost the entire vein (Fig. 6.15a). Biotite occurs as inclusions in quartz and in mutual contact with quartz, suggesting a contemporaneous crystallisation of the two mineral phases, but with biotite slightly precipitating earlier than quartz (Fig. 6.15b, c). Locally, replacement of biotite by chlorite is observed. The relationship between quartz and plagioclase suggests a cogenetic crystallisation as these two phases seem to be in textural equilibrium (Fig. 6.15d, e). These alteration minerals are associated with chalcopyrite that locally replaces pyrite. Both quartz and plagioclase are enclosed within chalcopyrite.

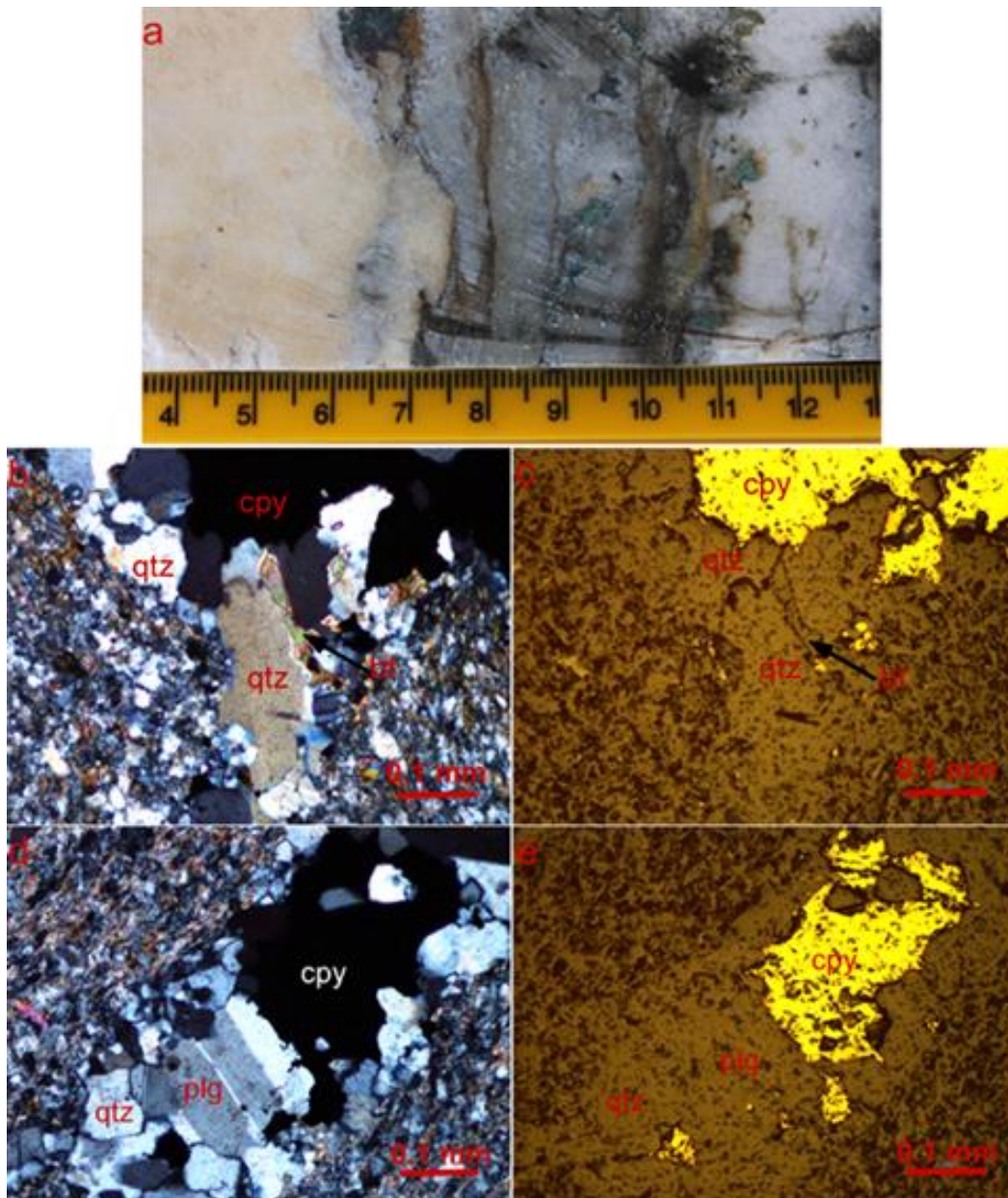


Fig. 6.15. (a), Photograph of massive zoned quartz-plagioclase-chalcopyrite-biotite vein (drill core HA 251-D; 659.40-659.78 m). (b), XPL transmitted light photomicrograph showing quartz (qtz)-biotite (bt)-chalcopyrite (cpy) association in the massive vein. (c), Reflected light photomicrograph of (b). (d), XPL transmitted light photomicrograph showing quartz (qtz)-plagioclase (plg)-chalcopyrite (cpy) association. (e), Reflected light photomicrograph of (d).

6.2.6. Calcite-sphalerite-galena $\pm$ chalcopyrite veins

These veins, which are very thin (~ 1 -2 mm) and not deformed, overprint the sodic-calcic alteration that is associated with the fine grained sandstone of the D'Kar Formation (see Fig. 6.1). The ore minerals that are associated with these veins include sphalerite, galena, and chalcopyrite. In places, sphalerite has irregular bleb-like inclusions of chalcopyrite

(i.e., chalcopyrite disease) (see Fig. 6.1b) and galena forms incomplete rims around the sphalerite. However, massive sphalerite crystals free of chalcopyrite disease are also observed. This massive sphalerite encloses both the chalcopyrite and the galena (Fig. 5.4g in Chapter 5).

Table 6.1
Summary of alteration types associated with the mineralisation of the Banana Zone Cu-Ag mineralisation

Alteration types	Mineral assemblages		Mode of occurrence		Associated ore minerals	
	Major	Minor	Major	Minor	Major	Minor
Sodic-calcic	alb, cal	chl, mt	Pervasive	none	none	
Phyllic	Qtz-ms-chl	alb	Pervasive		py, po, asp	cpy
Silicification	qtz	cal	Vein-controlled	Pervasive	py, cpy	cc
	qtz	bt, plg, chl	Vein-controlled	none	cpy	py
	qtz	ms	Vein-controlled	none	bn, cpy	
Propylitic	chl-cal-alb	ms, anh	Pervasive	Vein-controlled	cpy, bn	cc
Saussuritisation	chl-ms-cal		Pervasive	Vein-controlled	cpy,bn,cc	none
Carbonate	cal-qtz	ms, chl, alb, bt	Pervasive	Vein-controlled	cpy, bn, cc	hm, mt
	cal	sid	Vein-controlled	none	spec	none
						cv, bn,
Supergene	goe/lim-hm	none	Pervasive	none	cc, dg,	mal

alb, albite; bt, biotite; cal, calcite; chl, chlorite; goe, goethite; lim, limonite; hm, hematite; ms, muscovite; plg, plagioclase; sau, saussurite; sid, siderite; qtz, quartz; cc, chalcocite; cpy, chalcopyrite; dg, digenite; gn, galena; py, pyrite; po, pyrrhotite; asp, arsenopyrite; bn, bornite; sph, sphalerite; mt, magnetite; mal, malachite; spec, specularite; cv, covellite

CHAPTER 7

DISCUSSION

This chapter aims at interpreting and discussing the results obtained in this study, focusing on the petrographic and geochemical characteristics and their implications for sandstone classification, provenance, and source rocks characteristics, and the weathering intensity of the source area as well as tectonic setting of the region. It also covers the paragenetic sequence of ore and hydrothermal alteration minerals in the Banana Zone, fluid characteristics, ore deposition stages, as well as proposing a tentative genetic model of the Banana Zone mineralisation.

7.1. Sandstone classification

Sandstone classification followed in this study is based on the matrix content and mineralogical composition (**Dott, 1964; Pettijohn et al., 1973, 1987; Folk, 1974, 1980**). Given that the matrix content of the investigated sandstone is ≤ 15 vol.% modal and its Al_2O_3 contents is relatively low (average 8.48 and 9.17 wt.% in the Ngwako Pan and D'Kar Formations, respectively), the sandstone of both the Ngwako Pan and D'Kar Formations is classified as arenite. The relative proportion of the three major framework grains, namely monocrystalline quartz (Qm), total feldspar (F), and total lithic fragments (Lt), suggests that the sandstone of the Ngwako Pan Formation can be further classified as arkose, whereas the sandstone of the D'Kar Formation is classified as sub-arkose, and arkose (Fig. 3.2). Based on the geochemical classifications proposed by **Herron (1988)** sandstone from both D'Kar and Ngwako Pan Formations is classified as arkose (Fig. 7.1a), (see also **Lindsey, 1999**). This is further supported by the high content of Ba, Sr, and Rb signifying the enrichment in feldspar. Therefore, the investigated sandstone can be classified as potassic sandstone following the classification of **Blatt et al. (1972)** (Fig. 7.1b).

In addition, sandstone of the Ngwako Pan and D'Kar Formations is poorly to moderately-sorted (as indicated by variation in size of the grains) and the roundness ratio between fine- and coarse-grained is correlated with the size of the clasts pointing towards texturally

immature to submature sediments (Folk, 1951). Mineralogical maturity can be deduced from $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios that reflect the abundance of quartz, feldspar, and clay contents within the sediments (Potter, 1978). Low average values of $\text{SiO}_2/\text{Al}_2\text{O}_3$ found in the studied samples (8.97 for the sandstones of the Ngwako Pan Formation and 8.05 for the sandstones of the D'Kar Formation) are indicative of low maturity (Table 4.1). The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios can also be used in determining chemical maturity as the $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio reflects the liable nature of plagioclase relative to K-feldspar (Pettijohn et al., 1987). The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ average values of 0.69 for the sandstone of the Ngwako Pan Formation and 0.80 for the sandstones of the D'Kar Formation indicate that the samples are chemically immature (Table 4.1).

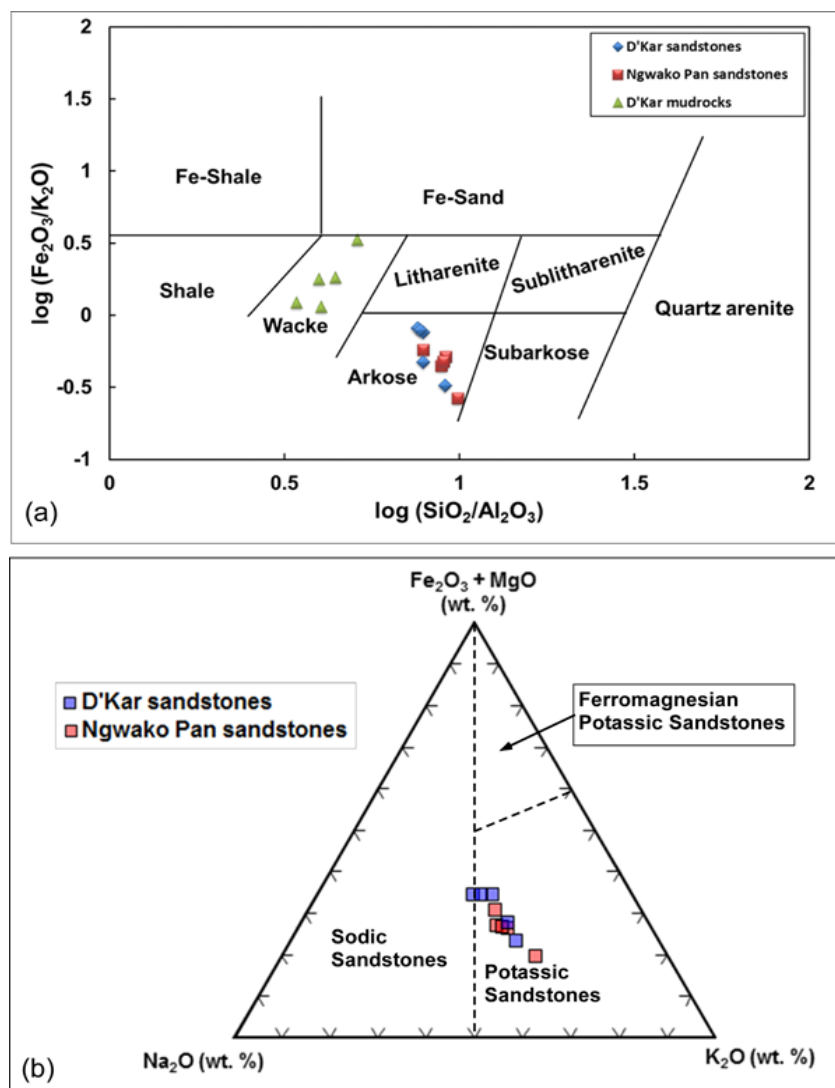


Fig. 7.1: Chemical classification of the investigated sandstones based on the Herron (1988) diagram. (b) Na_2O - K_2O - $(\text{Fe}_2\text{O}_3+\text{MgO})$ ternary diagram showing the chemical composition of the sandstones from the D'Kar and Ngwako Formations (after Blatt et al., 1972).

7.2. Provenance and source rock characteristics

Petrographic observations of the sandstone of both the Ngwako Pan and the D'Kar Formations point to a felsic igneous source rock with minor contributions from metamorphic or meta-sedimentary and sedimentary sources. A felsic volcanic source is supported by the high amount of monocrystalline inclusions-free quartz with straight extinction (for both the Ngwako Pan and the D'Kar Formations) and the presence of volcanic rock fragments (Fig. 3.4a) in the sandstones of the Ngwako Pan Formation. Furthermore, sandstone of the Ngwako Pan Formation contains myrmekite and granitic clasts (Fig. 3.4b, c), medium- to coarse-grained monocrystalline quartz containing microfracture-controlled fluid inclusions and exhibiting both undulose and non-undulose extinction, polycrystalline quartz composed of fewer crystals of more equant shape with little or no intercrystalline suturing (Fig. 3.3a). These characteristics, together with the presence of micropertite, point to a possible plutonic source (**Blatt, 1992**).

This felsic igneous provenance is further supported by major elements and their ratios, particularly Al_2O_3 and TiO_2 . Many studies in tropical regions have shown that Al and Ti do not fractionate during sedimentary processes and the $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios in the sediments can be considered as a pristine geochemical signature of the source rock composition (e.g., **McLennan et al., 1983; Taylor and McLennan, 1985; Schieber, 1992; Garcia et al., 1994; Sugitani et al., 1996; Huntsman-Mapila et al., 2005; Paikaray et al., 2008; Spalletti et al., 2012; Absar and Sreenivas, 2015; Zhou et al., 2015**). $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio normally varies from 3-8 in mafic rocks, 8-21 in intermediate rocks and 21-70 in felsic rocks (**Sugitani et al., 1996, 2006; Hayashi et al., 1997**). The $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio in the investigated samples shows values of 24.46-37.05, 21.36-29.29 and 18.55-21.32 for the sandstone of the Ngwako Pan Formation, the D'Kar Formation and mudrocks of the D'Kar Formation, respectively. The $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio indicates that the sandstone of both the Ngwako Pan and the D'Kar Formations was likely derived from felsic igneous rocks, whereas the chemical composition of the mudrocks of the D'Kar Formation is consistent with the contribution of an intermediate source rock. In addition, the result obtained from the TiO_2 -Zr plot (Fig. 7.2) shows that the source material for the sandstones of both formations is of felsic origin, whereas for the mudrocks it is of intermediate origin (**Schreiber et al., 1992**).

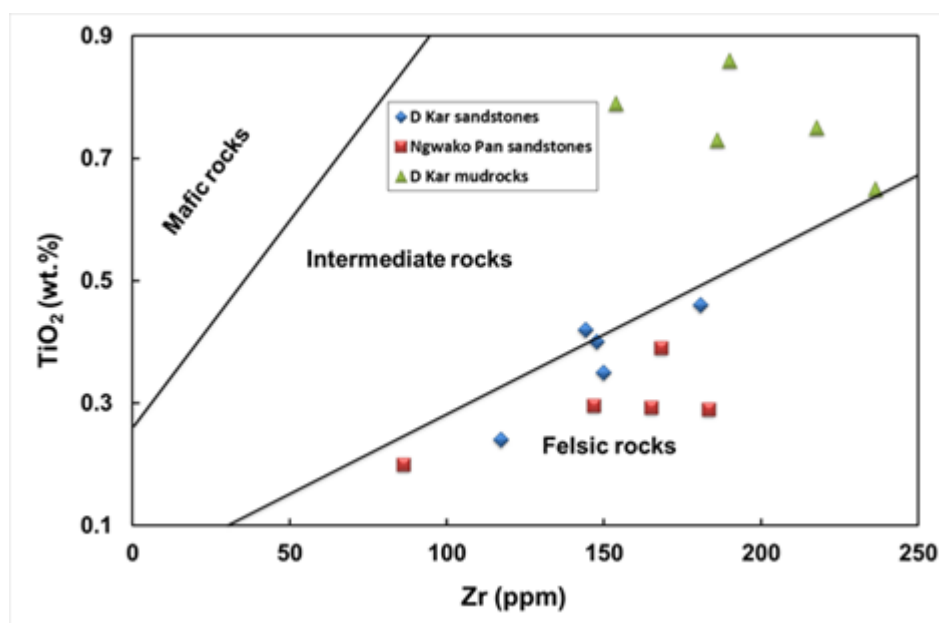


Fig. 7.2: TiO_2 vs Zr plot for the samples (after Hayashi et al., 1997).

The relative abundances of Cr, Co, Ni, and V in siliciclastic sediments are useful in distinguishing mafic from felsic sources, as these elements are preferentially concentrated in mafic than felsic rocks (Cullers, 1995; Huntsman-Mapila et al., 2005; Nagarajan et al., 2007; Armstrong-Altrin et al., 2013; 2014; 2015). The sandstone of both the Ngwako Pan and the D'Kar Formations has low concentrations of Cr, Co, Ni, and V, demonstrating a felsic provenance. On the other hand, the mudrocks of the D'Kar Formation have higher concentrations of Cr, Co, Ni, and V than the sandstone of both the Ngwako Pan and the D'Kar Formations (Table 4.1), but these values are lower than the sediments derived from mafic rocks (see Cullers, 1995). The Cr, Co, Ni, and V values demonstrated by the mudrocks from the D'Kar Formation suggest a mixed source. In addition, the distribution of trace elements, such as Ni and Cr may help further constraining the location of the source rocks. Taylor and McLennan (1985) suggested that Cr and Ni distributions such as those of Ngwako Pan and the D'Kar sandstone (Fig. 7.3) are consistent with a post-Archean source. This is further supported by $\text{Gd}_\text{N}/\text{Yb}_\text{N}$ ratios that are less than 2 (Table 4.2), thus suggesting that they were derived from less HREE-depleted Archean and/or post-Archean sources (Armstrong-Altrin et al., 2004). Similarly, the La/Co, Th/Co, Th/Cr, and Cr/Th ratios in felsic and mafic rocks show significant variations and as such are more useful in constraining the provenance of siliciclastic rocks (Wronkiewicz and Condie, 1990; Cox et al., 1995; Cullers, 1995). These ratios for the sandstones of both the Ngwako Pan and the D'Kar Formations are

comparable to the range of sediments derived from felsic source rocks, whilst the mudrocks of the D’Kar Formation are comparable to sediments derived from mixed source rocks (Table 4.3).

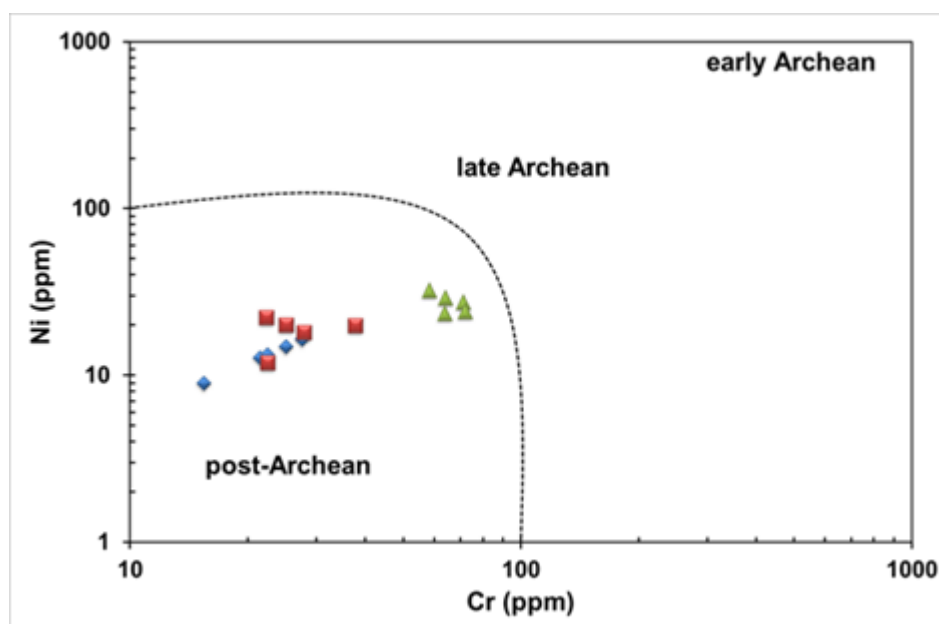


Fig. 7.3: Ni-Cr cross plot of major oxides and trace elements of the sandstones from the D’Kar and Ngwako Pan Formations and mudrocks of the D’Kar Formation. (Field boundaries after **Taylor and McLennan, 1985**).

Chondrite-normalised REE patterns and the size of the Eu anomaly have proven their utility as proxies for the assessment of the provenance of siliciclastic sediments (**Taylor and McLennan, 1985; McLennan, 1989; Cullers and Podkovyrov, 2000; Singh, 2009; Shynu et al., 2011, Etemad-Saeed et al., 2011**). Low LREE/HREE ratios and absence of Eu anomalies are characteristics of mafic rocks, whereas more felsic rocks are characterised by higher LREE/HREE ratios and negative Eu anomalies (**Cullers and Graf, 1984; Cullers, 1994, 2000**). Furthermore, LREE enrichment, negative Eu anomaly and almost flat HREE patterns are indicative of cratonic source rock with the presence of felsic and meta-sedimentary components (**Gu et al., 2002; Das et al., 2006**). Therefore, high LREE/HREE ratios (Table 4.2), significant negative Eu anomalies, LREE enrichment, and almost flat HREE patterns as observed in the samples from both the Ngwako Pan and D’Kar Formations (Fig. 4.1a, b) point to a source area of felsic composition. Moreover, chondritic Y/Ho ratios (Table 4.2) are typical for terrigenous materials derived from a magmatic source (e.g., **Klinkhammer et al., 1994; Bau and**

Dulski, 1999). Accordingly, the average Y/Ho values of the studied samples are consistently below 30 supporting their derivation from a magmatic source.

To further underpin the provenance identification, the average REE data for pre-Neoproterozoic rocks (granites, gneisses, quartz diorite and schist) from Mesoproterozoic Kibara Belt (KIB) and the Karagwe-Ankole Belt in Central Africa (**Debruyne et al., 2015**) were compared with the average REE data of the investigated samples (Fig. 7.4). The general shapes of the REE pattern of the investigated samples are comparable to the average D1-granite, D1-gneisses, and D2-granites. This comparison suggests that the sediments for the studied samples were derived largely from felsic rocks.

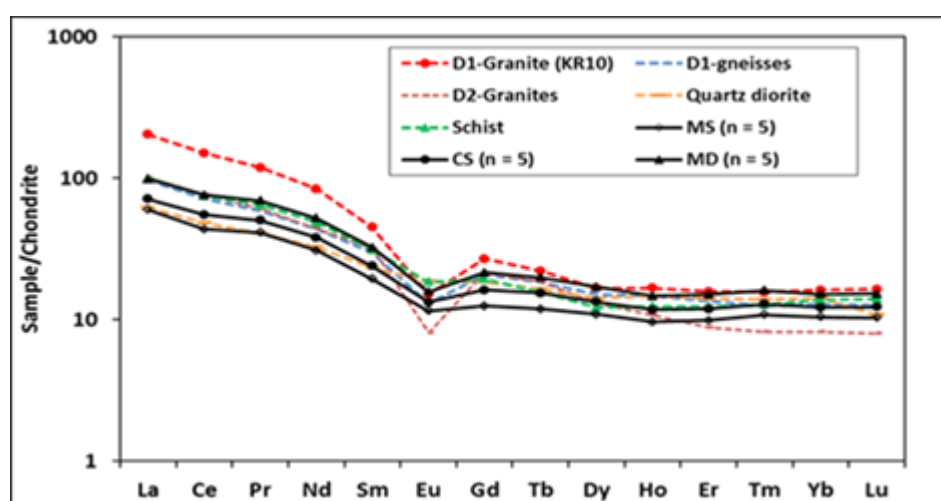


Fig. 7.4: Average chondrite-normalised REE patterns for Ngwako Pan and D'Kar sandstones, and D'Kar mudrocks compared to average chondrite-normalised REE patterns of pre-Neoproterozoic rocks (granites, gneisses, quartz diorite and schist) from different belts in Central Africa compiled from **Debruyne et al., 2015**. Normalisation values are from **Taylor and McLennan, 1985**. n = number of samples.

However, recrystallised quartz (Fig. 3.3d), polycrystalline quartz grains having polygonal crystal boundaries at 120° angle (Fig. 3.3c), and the presence of metamorphic rock fragments (Fig. 3.4d) in the sandstone also reflect a small contribution from a metamorphic source rocks. High amount of mudstone clasts (Fig. 3.4e) and the presence of microcrystalline chert grains (Fig. 3.4f) also suggest a contribution of a sedimentary source.

From the discussion elaborated above, both petrographic and geochemical data suggest that the sediments of the Ngwako Pan and the D'Kar Formations were largely derived from felsic igneous rocks with minor contribution from metamorphic and sedimentary rocks. Hence, considering the paleogeography of the area during the Neoproterozoic, the

probable source rocks are located in the Palaeoproterozoic basement rocks (granitoids and granitic gneiss) and the Mesoproterozoic Kgwebe volcanic rocks exposed north from the study area.

7.3 Weathering intensity of the source area

The intensity of chemical weathering in the source area is controlled by the climate and the rate of tectonic uplift (**Wronkiewicz and Condie, 1987**). Therefore, information regarding the tectonic activity and climatic conditions in the source area can be obtained from weathering indices of sedimentary rocks. Moreover, modal compositions of sand can provide information on the intensity of chemical weathering as well as paleoclimatic conditions (**Dickinson and Suczek, 1979; Van de Kamp, 2010**). As already demonstrated in the Qm-F-Lt diagram (Fig. 3.2), the sandstone from both formations is classified as arkose and subarkose; with K-feldspar more abundant than plagioclase. According to **Van de Kamp (2010)**, arkoses in which K-feldspar is dominant in abundance are formed in humid temperate and subtropical climates. It has to be noted that prolonged chemical weathering of feldspar-rich source rocks in the hot and humid environment will skew the modal composition of sandstones toward a quartz-bearing composition (e.g., **Garzanti et al., 2013; Schneider et al., 2016**).

As established by **Nesbitt and Young (1982)**, the relationships among alkalis and alkali-earth elements can be used as a measure of the intensity and duration of weathering. The mobility of these elements, especially during diagenesis can be evaluated by using the Chemical Index of Weathering (CIW; **Harnois, 1988**): $Al_2O_3 / (Al_2O_3 + CaO^* + Na_2O) \times 100$ (in molecular proportions), where CaO^* is the CaO content within the silicate fraction only. The CaO detected in the investigated rock samples is poorly correlated ($r = 0.23$, for the sandstone of the Ngwako Pan Formation) to negatively correlated ($r = -0.91$, $r = -0.93$, for the sandstone and mudrocks of the D'Kar Formation, respectively) with Al_2O_3 implying that CaO is probably not incorporated in the silicate phases, but coming from calcite cements (see Chapter 3). Moreover, the investigated samples have highly variable CaO contents due to variation in calcite abundance and this may produce misleading conclusions if the CIW are used to infer the intensity of weathering (**Cullers, 2000**). To avoid the bias due to Ca distribution within the cements, **Cullers (2000)** proposed a modified Chemical Index of Weathering (CIW'), in which CaO is not considered: $Al_2O_3 /$

$(\text{Al}_2\text{O}_3 + \text{Na}_2\text{O}) \times 100$. The CIW' values of the investigated samples vary from 79 to 81 (average 79) for the sandstone of the Ngwako Pan Formation, from 77 to 80 (average 79) for the sandstone of the D'Kar Formation and from 81 to 89 (average 86) for the mudrocks of the D'Kar Formation. According to **Cullers (2000)** these high CIW' values are indicative of intense chemical weathering of the source rock, which may reflect a change of climate towards warm and humid conditions in the source area (**Jacobson et al., 2003**). This therefore suggests that source rocks of the sediments now found at the Banana Zone underwent intense chemical weathering prior to their deposition.

7.4. Tectonic setting

The chemical and petrographic composition of the sandstone can also reflect the tectonic setting of the source rock during the weathering and erosion (**Dickinson, 1985**). The Qt-F-L and Qm-F-Lt diagrams (Fig. 7.5a, b) indicate that the sandstones were probably derived from a continental block. As indicated by **Maynard (1984)** and **Blatt (1992)**, the enrichment of quartzo-feldspathic detritus coupled with high K-feldspar/plagioclase ratio in the sandstones is indicative of sands derived from rifted areas associated with continental separations.

During the past 25 years, major elements patterns of clastic rocks were used to discriminate the tectonic settings of the ancient siliciclastic depositional systems. The discrimination was based on the $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio and the SiO_2 content, as they vary consistently in different tectonic environments (**Maynard et al., 1982; Roser and Korsch, 1986**). However, the use of these discrimination diagrams has recently received massive criticism from numerous workers (e.g., **Armstrong-Altrin and Verma, 2005; Weltje, 2006; Ryan and Williams, 2007; von Eynatten et al., 2012; Armstrong-Altrin, 2015**) as their performance is unsatisfactory. Therefore, in this study, new discriminant-function-based multi-dimensional diagrams recently introduced by **Verma and Armstrong-Altrin (2013)** were used to decipher the tectonic setting of the investigated samples. These diagrams, which discriminate three tectonic settings, viz, arc (active volcanism), continental rift (extension), and continental collision (compression) were constructed from $\log_e$ ratio transformation of all major elements. In these diagrams major elements (SiO_2 to P_2O_5) are adjusted to 100 % on a volatile free basis (see Appendix A4) resulting in two

multi-dimensional diagrams for high-silica ($\text{SiO}_{2\text{adj}}=63\%–95\%$) and low-silica ($\text{SiO}_{2\text{adj}}=35\%–63\%$) clastic sediments. Since the adjusted SiO_2 concentration is greater than 63 %, all the investigated samples from both formations plots in the high-silica diagram (Fig. 7.5c). In this diagram, the majority of the sandstones of both the Ngwako Pan and the D’Kar Formations plot in the field for a rift setting suggesting a passive margin (cf. **Verma and Armstrong-Altrin, 2016**), except for two samples (one sample each) from both the Ngwako Pan and the D’Kar Formations that lie in the collision field, close to the boundary between the collision and the rift field. This result is consistent with the result inferred from the Q-F-L ternary diagrams (Fig. 7.5a, b) in which the sandstone of both the Ngwako Pan and the D’Kar Formations plots in the continental block (transitional continental) field. Moreover, a continental rift setting inferred from the high silica diagram is consistent with the geology of the area (**Modie, 1996**). On the other hand, in the high silica diagram (Fig. 7.5c), the mudrocks of the D’Kar Formation plot in the field for arc setting suggesting an active margin (cf. **Verma and Armstrong-Altrin, 2016**).

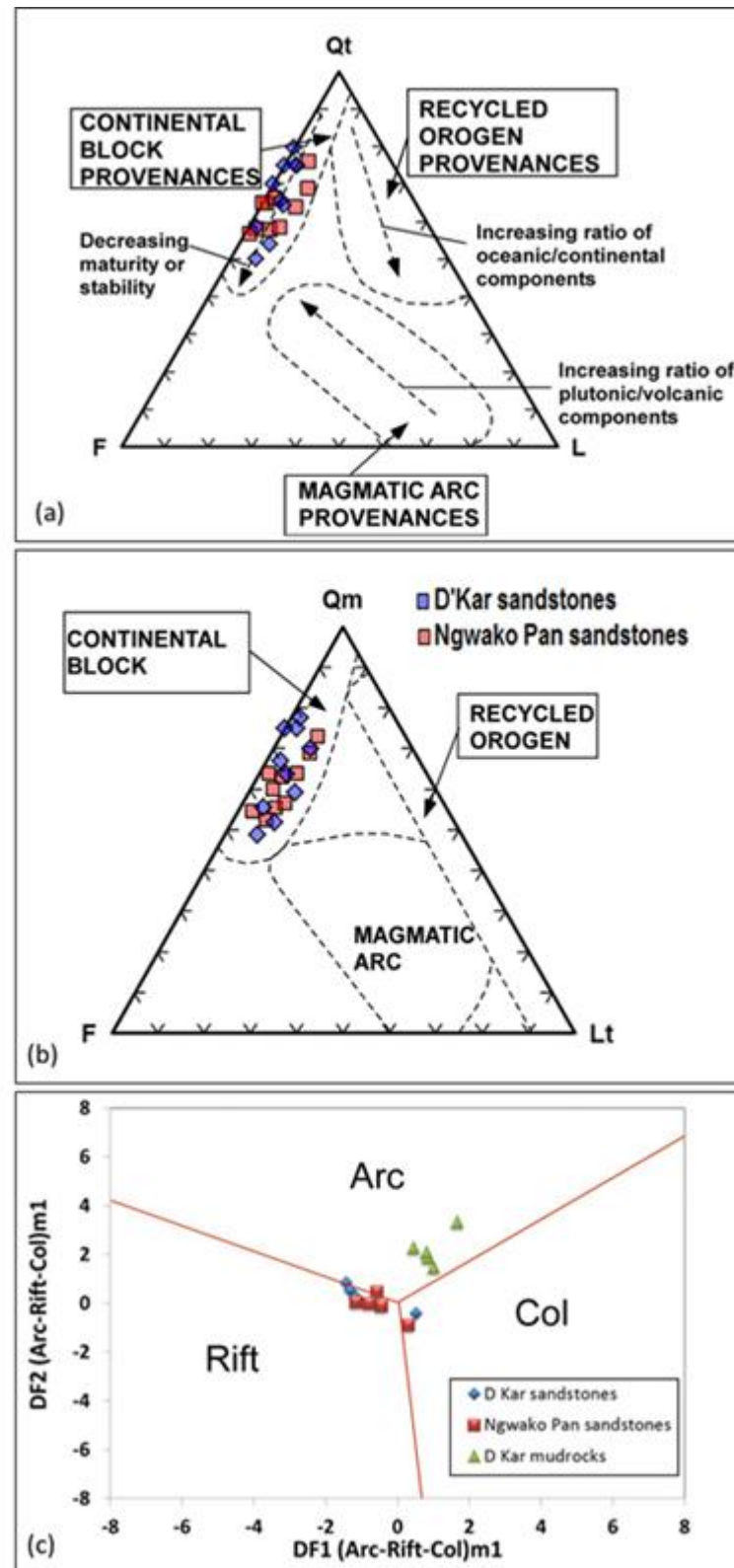


Fig. 7.5: (a, b) Qt-F-L and Qm-F-Lt ternary plots for palaeotectonic provenance of the studied sandstones (after Dickinson, 1985). Qt=Total quartz grains (monocrystalline + polycrystalline), F=Total feldspar grains (plagioclase + K-feldspar), L=Unstable siliciclastic lithic fragments. (c) Discriminant function multi-dimensional diagram for high-silica clastic sediments (Verma and Armstrong-Altrin, 2013). The discriminant functions DF1 and DF2 equations for high-silica clastic sediments are:

$$\begin{aligned}
DF1_{(Arc-Rift-Col)m1} = & (-0.263 \times \ln (TiO_2/SiO_2)_{adj}) + (0.604 \times \ln (Al_2O_3/SiO_2)_{adj}) + (-1.725 \times \ln (Fe_2O_3t/SiO_2)_{adj}) \\
& + (0.660 \times \ln (MnO/SiO_2)_{adj}) + (2.191 \times \ln (MgO/SiO_2)_{adj}) + (0.144 \times \ln (CaO/SiO_2)_{adj}) + (-1.304 \times \ln \\
& (Na_2O/SiO_2)_{adj}) + (0.054 \times \ln (K_2O/SiO_2)_{adj}) + (-0.330 \times \ln (P_2O_5/SiO_2)_{adj}) + 1.588; DF2_{(Arc-Rift-Col)m1} = (- \\
& 1.196 \times \ln (TiO_2/SiO_2)_{adj}) + (1.064 \times \ln (Al_2O_3/SiO_2)_{adj}) + (0.303 \times \ln (Fe_2O_3t/SiO_2)_{adj}) + (0.436 \times \ln \\
& (MnO/SiO_2)_{adj}) + (0.838 \times \ln (MgO/SiO_2)_{adj}) + (-0.407 \times \ln (CaO/SiO_2)_{adj}) + (1.021 \times \ln (Na_2O/SiO_2)_{adj}) + \\
& (-1.706 \times \ln (K_2O/SiO_2)_{adj}) + (-0.126 \times \ln (P_2O_5/SiO_2)_{adj}) - 1.068.
\end{aligned}$$

7.5. Paragenetic sequence

Crosscutting relationships, inclusion and replacement relationships demonstrate that sodic-calcic (albitisation) alteration is the earliest formed alteration type in the Banana Zone Cu-Ag mineralisation area. The early timing of this alteration is documented by the replacement of plagioclase (albite) by fine-grained muscovite (as observed petrographically) and the crosscutting of the albitisation by calcite veinlets (see Fig. 6.1a). It has to be noted that the hydrothermal origin of this albite can be debatable because of the similarity in shape of both the detrital albite and hydrothermal albite and the lack of spatial association of albite and ore minerals. The inclusions of sericite/muscovite in other gangue minerals and the overprinting of the phyllic alteration by other types of alteration, except for sodic-calcic alteration demonstrate that phyllic alteration is subsequent to sodic-calcic alteration. This gradual change of sodic-calcic alteration to phyllic alteration is marked by an increase in the proportion of muscovite probably due to the influx of K-rich fluids in the system. In the sericitisation process, silica is produced as a by-product (**Evans, 1993; Guilbert and Park, 1996**) and the resultant silica may be precipitated as secondary silica that increases the proportion of quartz in the altered rock or manifested as quartz veins. The silicification is accompanied by the dissolution of the carbonate minerals creating voids due to the high acidity of fluids involved in the silicification process (**Gorrepati et al., 2010**). A subsequent decrease in H^+ activity and fluid temperature (as a result of fluid-rock interaction) will favour the stability of propylitic alteration mineral assemblage (see Fig. 7.6). The chlorite from Banana Zone propylitic assemblage is Mg-rich (see Appendix A2), suggesting the involvement of seawater (**Hayes, 1970**), typical of evaporitic environments (**Selley et al., 2005**). Sodic-calcic, phyllic, and propylitic alterations are overprinted by carbonate-quartz alteration as demonstrated by the pervasive

overprint of the earlier formed hydrothermal minerals by calcite (see Fig. 6.6), inclusions of sericite/muscovite, chlorite and quartz within the calcite and crosscutting and displacement of sericite layers by a calcite-quartz veinlet. Textures of the calcite cement (e.g., fine-grained, untwinned calcite and coarse-grained, twinned calcite) demonstrate that there was either more than one calcite event or that the calcite cementation occurred over an extended period of time as fluid composition evolved (**Sutton and Maynard, 2005**).

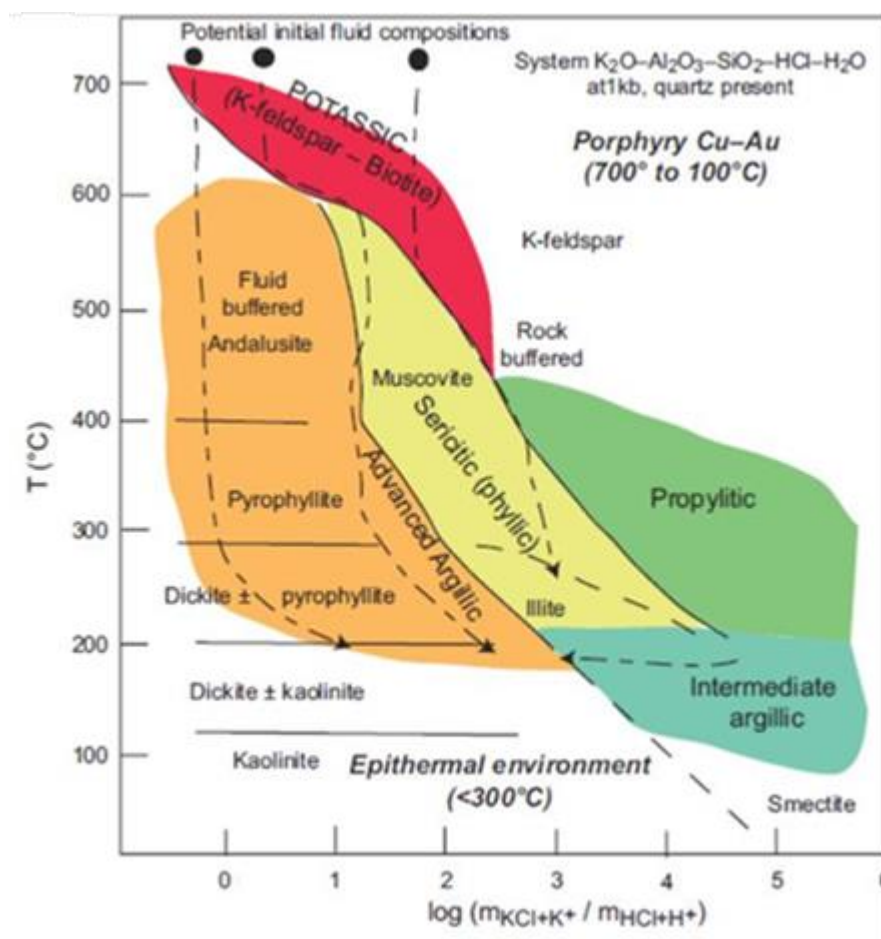


Fig. 7.6: Illustration of alteration type as function of temperature, K^+ and H^+ activities (after **Simon et al., 2005**).

Given that alteration mineral assemblages are associated with ore minerals and no ore minerals were observed associated with sodic-calcic alteration, it is therefore reasonable to postulate that the pyrite-pyrrhotite-chalcopyrite assemblage associated with the phyllic alteration is the earliest ore mineral association. In this assemblage, pyrite is enclosed in both pyrrhotite and chalcopyrite, thus allowing the identification of pyrite as the earliest mineral. Subsequent to pyrite-pyrrhotite-chalcopyrite assemblage is the sphalerite-galena-

chalcopyrite; chalcopyrite-bornite and bornite-chalcocite associations hosted in the propylitised and carbonatised units. These assemblages represent the main mineralisation stage that formed at a wide range of temperature and fluid composition (see Fig. 7.6). Textural evidence indicates a contemporaneous crystallisation between chalcopyrite and bornite, and bornite and chalcocite (see chapter 6). An association of chalcopyrite-bornite-chalcocite with carbonate-rich zones that overprint other alteration types demonstrates that this ore mineral association is of later origin. Locally, in calcite-quartz veins that crosscut the Ngwako Pan Formation sandstone, specular hematite and magnetite are observed, with hematite replacing magnetite.

Crosscutting relationships between veins were not observable both at macroscopic and microscopic scale, making it very challenging to establish a chronological sequence of the veins. However, the deformation features observed in these veins are used here to come up with their paragenetic sequence. This could be more complex than what is demonstrated here. As mentioned earlier, the veins are either parallel or oblique to the stratification. But irrespective of their mode of emplacement, these veins are either deformed or undeformed and they consist of sulphides or iron-oxides, demonstrating successive vein generations. The occurrence of either sulphide or iron-oxides in the veins most likely reflects a change in the redox state of the mineralising fluids.

Though in some of the deformed veins gangue minerals are deformed, the accompanying sulphides are less noticeably deformed. In addition, in tectonically brecciated veins sulphides occur as breccia cement. The presence of undeformed sulphides in deformed veins and the occurrence of sulphides as cement in breccias indicate the late exploitation of weakens planes in the veins by the mineralising fluids. Therefore, on the basis of deformation status, veins are discriminated into either pre- to syn-tectonic (deformed veins) or post-tectonic (undeformed veins). Considering that the deformation features were not observed in the sulphide mineralisation in the deformed veins, the mineralisation could be post-tectonic, but it must be noted that different sulphides behave differently to deformation stress (i.e., brittle or ductile manner).

Supergene processes redistributed minerals after the major hypogene/hydrothermal alteration events. This is demonstrated by the association of chalcocite with iron oxides/hydroxides that indicates formation during supergene enrichment (**Van Langedonck et al., 2013**). This is further demonstrated by chalcocite grains that are

associated with secondary porosity and lack of intergrowths with hypogene bornite or chalcopyrite (**cf. Torremans et al., 2012 and references therein**). Furthermore, the replacement of hypogene ore minerals along boundaries by chalcocite (rim pattern) (see Fig. 6.3) reveals a supergene influence (**Craig and Vaughan, 1994; Sillitoe, 2005; Van Langendonck et al., 2013**). Covellitisation of primary copper minerals is also observed in this zone of secondary enrichment. Covellite replaces chalcocite along the cleavage planes. Significant quantities of malachite are also observed.

Based on the discussion elaborated on above, the paragenetic sequence reported in Fig. 7.7 was proposed.

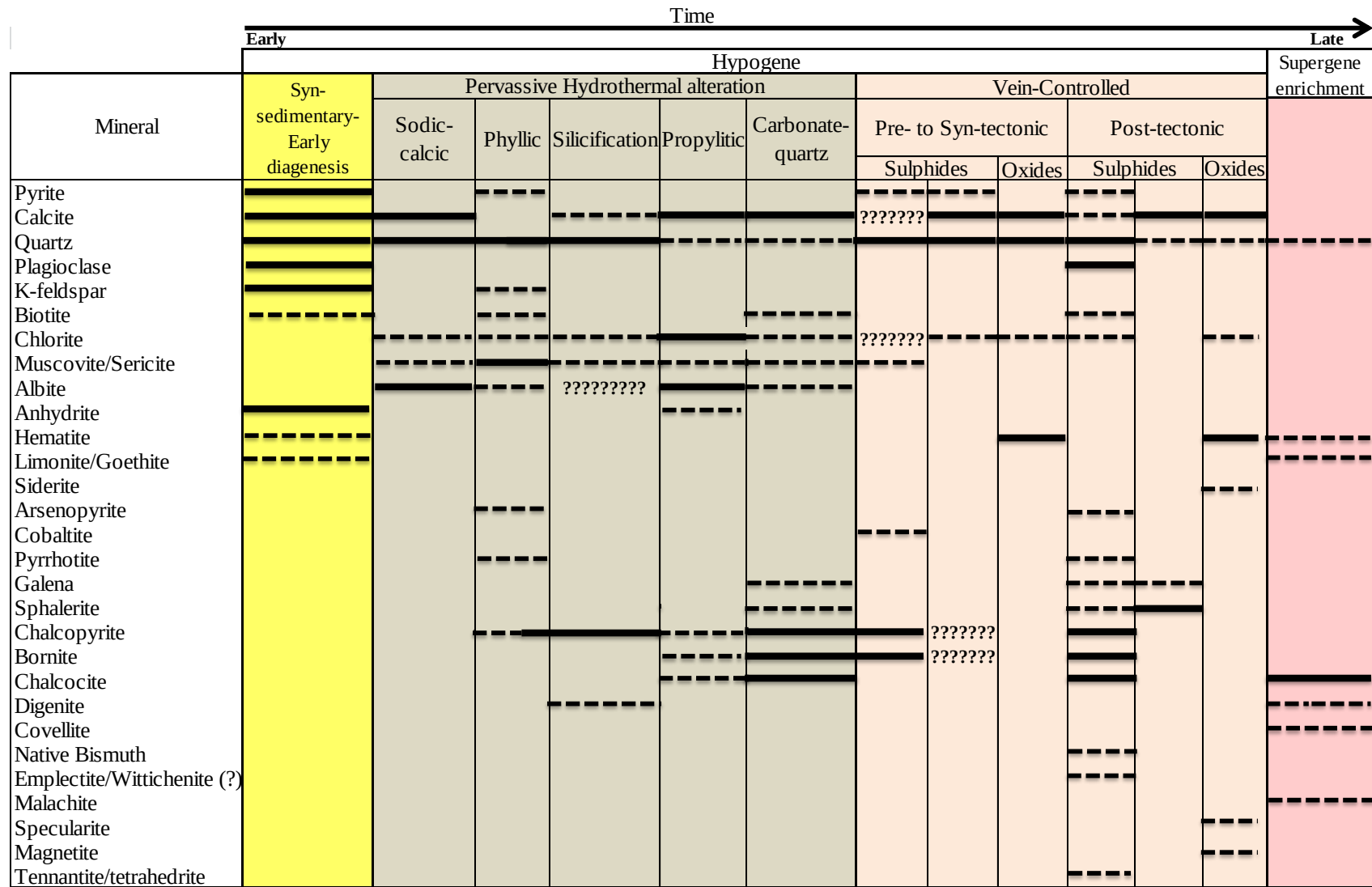


Fig. 7.7: Paragenetic sequence of the Banana Zone Cu-Ag mineralisation. The thicker the line the more abundant the mineral

7.6. Characteristics of mineralising fluids

Wall-rock alteration analysis is an important tool in assisting deciphering mineralising fluid features such as:

1. Temperature
2. Anion-cation composition
3. pH and redox state
4. mineralising fluid evolution

So, by studying the alteration patterns, one can then unravel the physicochemical composition of the hydrothermal fluids that transported and deposited the associated ore minerals (see Fig. 7.6). It must be noted that without fluid inclusion data the exact P-T conditions cannot be robustly constrained.

From the above discussions it is evident that mineralisation span is from early diagenesis to metamorphism. This prolonged history of mineralisation demonstrates the involvement of multiple fluids of different origin and most probable different composition. This contention is further supported by a wide range of sulphur isotopic composition of sulphides, which range from -43.1 to + 3.5‰ (relative to V-CDT), obtained in the similar study conducted by **Hall (2013)** from the adjacent prospect area at Boseto Mine. In addition, the presence of hematite coating in detrital grains provides circumstantial evidence of oxidised fluids (see **Brown, 2009**). This is further demonstrated by the occurrence of bladed hematite in quartz-carbonate veins.

7.6.1. Temperature of the mineralising fluids

The mineralising fluids that led to the Banana Zone Cu-Ag occurrence certainly had a wide range of temperature ($\leq 60 - 400^{\circ}\text{C}$) as demonstrated by the following:

- a) the occurrence of the framboidal pyrite, which is reputed to precipitate at temperature between 20 and 60°C (cf. **Bartholome, 1974; Fleischer et al., 1976; Sweeney et al., 1986, 1991; Okitaudji, 1992, 2001; Machel, 2001; Hitzman et al., 2005; Dewaele et al., 2006**).

- b) The replacement of framboidal pyrite by Cu-sulphides (e.g. chalcopyrite). **Cowper and Rickard (1989)** have alluded that this replacement reaction is only possible at elevated temperature (but not exactly specified). However, **Quigley et al. (1987)** suggested a maximum temperature of around 130°C to have affected the Kupferschiefer based on the coal petrological results.
- c) Exsolution of chalcopyrite lamellae from bornite (as observed in Fig. 5.2f), that occurs upon heating around 200-250°C (**Durazzo and Taylor, 1982**).
- d) The myrmekitic intergrowth of chalcocite and bornite (Fig. 5.4e). In fact, below *ca.* 200°C chalcocite will exsolve bornite (**Yund and Kullerud, 1966; Sugaki et al., 1975; Hatert, 2005**).
- e) The occurrence of supergene chalcocite (low chalcocite; **Ramdohr, 1969, 1980**) that occurs from the transformation of hypogene chalcocite (hexagonal chalcocite) at 103.5°C (**Ramdohr, 1980; Fleet, 2006**) or formed from supergene solutions (**Ramdohr, 1969**).
- f) The relationship between Cu-sulphides and Mg-metasomatism. The predominance of Mg-chlorite over biotite demonstrates temperature ranging between 150° and 400°C (see **Seedorff et al., 2005**).
- g) The occurrence of deformed mineralisation within deformed veins and the intergrowth of sulphides with metamorphic minerals (e.g., rutile, see Fig. 5.4f) suggest either metamorphism of pre-existing mineralisation or precipitation of mineralisation at higher temperatures (~250°- 400°C) (**Selley et al., 2005**). This contention is further supported by the precipitation of native bismuth instead of bismuthinite that is favoured by mainly high temperatures $\geq 400^{\circ}\text{C}$ (**Tooth et al., 2008; Bineli Betsi et al., 2016**).
- h) The extensive development of albitisation indicates higher temperature (~450°C).

Most fluid inclusion studies in the region have recorded low temperatures (>115-220°C and 270-385°C) than what is suggested in this study (see **Hall, 2013; Schmandt et al., 2013; El Desouky et al., 2010; Dewaele et al., 2006; Cailteux et al., 2005**). It must be noted that mineral assemblages can be approximate indicators of temperatures of deposition as minerals have a wide range of temperature of crystallisation. Therefore,

without fluid inclusion data the exact temperature conditions cannot be robustly constrained.

7.6.2. Cations and volatiles compositions of the fluids

The Banana Zone mineralising fluids were certainly enriched in the following cations.

- a) Na-rich fluids as demonstrated by widespread of albitisation.
- b) K-rich fluids as shown by the development of K-rich alteration (e.g., phyllic) that totally overprinted the host rocks.
- c) Mg-rich as illustrated by the extensive development of Mg-rich chlorite and the association of mineralisation with chlorite-rich zones.
- d) Ca- and CO₂-rich fluids as supported by the extensive development of carbonate alteration and the predominance of calcite mineral in veins.
- e) SiO₂-rich fluids as demonstrated by the silicification process and the abundance of quartz as a vein mineral.

7.6.3. Salinity of the fluids

The presence of evaporite in the sedimentary sequence (cf. **Modie, 1996b**; section 3.2 of this study) and the dissolution of this evaporite may contribute to the salinity of the fluids resulting in saline fluids. The involvement of saline fluids is supported by data from **Hall (2013)**, who obtained the salinity of 15-20 wt. % equiv. NaCl from a study conducted at the nearby Boseto Mine. The salinity of the fluids is further supported by the fact that the sediments have accumulated in an intracratonic basin, and basinal fluids are reputed to be saline (**Kesler et al., 1995; Kesler, 2005; Hitzman et al., 2010**).

7.6.4. pH of the fluids

The mineralising fluids from Banana Zone were characterised by a wide range of pH conditions that vary from strongly acidic to alkaline conditions as supported by the following:

- a) Moderate to strong acidic fluids (see Fig. 7.6) as documented by the association of mineralisation with the phyllic alteration and the bleaching effect as result of removal of iron oxides from the rock.
- b) Weakly acidic fluids (see Fig. 7.6) as supported by the association of mineralisation with the propylitic alteration. The occurrence of coarse muscovite ($2M_1$) requires a lower ratios of aK^+/aH^+ (**Thompson, 1998**)
- c) Neutral to alkaline fluids as demonstrated by the association of mineralisation with the carbonate alteration. Carbonatisation is known to be promoted by fluids characterised by high partial pressures of carbon dioxide and neutral to alkaline pH (**Robb, 2005**).

7.7. Ore deposition stages

The mineralogy and style of mineralisation of the Banana Zone are similar to the CAC and the Polish Kupferschiefer mineralisation described by **Alderton et al. (2016)**. It is likely that the mineralisation at the Banana Zone is the result of multiple mineralising events that span from syn-diagenetic to post-tectonic events. This is demonstrated by textural evidence (e.g., framboidal, exsolution, replacement, deformational, etc.), diverse alteration patterns (see Chapter 5) and wide range of temperature (400 to $\leq 60^\circ\text{C}$). This protracted mineralisation event is also supported by the Re-Os model age dates obtained by **Hall (2013)**, which span from 1012 ± 17 Ma to 459 ± 37 Ma. These ages are consistent with the depositional age of the Ghanzi Group, which is bracketed between 1047 ± 24 Ma (youngest detrital zircons; **Kampunzu et al., 2000**) and 530 Ma (the date for peak metamorphism; **Ahrendt et al., 1978**). The first mineralising event that includes framboidal pyrite followed, and sometimes replaced by fine-grained disseminated mineralisation and mineralisation filling in pore spaces, is an early diagenetic to diagenetic

typical stratiform mineralisation (**El Desouky et al., 2009; Ridley, 2015; Alderton et al., 2016**). The early diagenetic origin of this mineralisation event is supported by the occurrence of framboidal pyrite (see **Dewaele et al., 2006; Large et al., 2009, 2011**) and the negative $\delta^{34}\text{S}$ values obtained by **Hall (2013)** at Boseto Mine that indicate formation from bacteriogenic reduction of seawater sulphate. Micro-laminations in siltstone hosting euhedral pyrite (see Fig. 4.6b) are perhaps indicative of algal activity (**Hinchey, 2010**). The occurrence of Cu-sulphides as fine-grained disseminations that locally display a framboidal texture and their distribution along the stratification (Fig. 5.2a) support the diagenetic origin of this mineralisation (**Brems et al., 2009** and references therein). Framboidal Cu-sulphides indicate a formation by replacement of framboidal Fe sulphides, and this replacement reaction is only possible at an elevated temperature (**Cowper and Rickard, 1989**). The presence of Cu-sulphide minerals in intergranular micro pores and their association with authigenic quartz indicate that the mineralisation is paragenetically late. Based on these observations the copper mineralisation postdates compaction and cementation by authigenic quartz and calcite (see **El Desouky et al., 2008a, b**). The sulphide paragenesis observed in the Banana Zone is comparable to the sulphide paragenesis documented in similar sediment-hosted deposits in the CAC (see **El Desouky et al., 2008a**, and b).

The second mineralisation event includes sulphide in deformed veins, porphyroblast and boudinage sulphides, brittle deformed sulphides, cleavage-controlled mineralisation (both in cleavage of rocks and cleavage of minerals) and breccia cements and is considered as pre- to syn-tectonic in origin. These veins and breccias crosscut the stratification, suggesting that this mineralisation event is paragenetically later than the first one and may indicate later remobilisation of metals. This is further supported by the overlapping of $\delta^{34}\text{S}$ values obtained from vein- and shear-hosted sulphides and disseminated sulphides (**Hall, 2013**), suggesting that structurally controlled sulphides were derived from dissolution of disseminated sulphides during deformation. The similarity in mineralogy and textures characterising both disseminated and accompanying veins is also indicative of remobilisation of early sulphides. However, this same feature may suggest simultaneous formation (cf. **Sillitoe et al., 2010**). The crosscutting relationship between vein sets further testifies to the multiple mineralisation events within the Banana Zone. The presence of biotite in these veins implies emplacement during metamorphism (cf. **Torremans et al., 2013**). The presence of mineralisation in fracture zones (i.e., veins, veinlets and breccias)

and along cleavages highlights the importance of the structural control of fluid flow during mineralisation.

A younger mineralisation event has been observed in veins that are not affected by deformation. These veins crosscut both the sandstone of the Ngwako Pan Formation and the siltstones and shales of the D'Kar Formation. Both the gangue and ore minerals in these veins seem to be not deformed and/or less deformed, signifying their emplacement late and/or after the deformation and metamorphism, hence, the mineralisation is likely late- to post-tectonic.

7.8. Tentative Genetic Model

Denudation of the basement units (i.e., the Palaeoproterozoic granitoids and gneisses and the Mesoproterozoic Kgwebe Volcanics) provided the sediments source (see section 7.2 above) of the Neoproterozoic sedimentary sequence hosting the Banana Zone Cu-Ag mineralisation. These sediments in addition to the main framework clasts (i.e., quartz and feldspar), also contain accessory biotite, amphibole, pyroxene, Fe-(Ti) oxides and rutile. As noted by **Walker (1989)**, **Brown (1992)**, and **Hitzman et al. (2005)**, these minerals contain traces amounts of Cu (> 50 ppm), which may represent Cu mineralisation sources upon leaching. Additionally, the metabasalts of the Kgwebe Volcanics may constitute the primary Cu source rocks (**cf. Borg, 1991**). Sedimentation of the Ngwako Pan and D'Kar Formations took place within an intracratonic rift basin (**Modie, 1996b; Kelepile et al., 2017**), with the Ngwako Pan Formation either originally deposited under oxidising conditions or was rapidly oxidised during burial and diagenesis (as supported by presence of hematite-rich components; **cf. Hitzman et al., 2010**) and the D'Kar Formation deposited under reduced conditions (as supported by the presence of organic-rich components; **cf. Cailteux et al., 2005; Hitzman et al., 2010**). Though there is no evidence of evaporitic units in the Ngwako Pan Formation as drilling stopped few metres after intersecting this formation, presumably, the evaporites are sitting at deeper levels as suggested by **Modie (1996b; see section 2.2. of Chapter 2)** and as observed by **Borg and Maiden (1989)** at Klein Aub area, Namibia. Likewise, there is local occurrence of marine evaporites in the D'Kar Formation as revealed by petrographic investigations (see section 3.2 of Chapter 3).

Subsequent burial and compaction during early diagenesis may have triggered significant fluid expulsion (connate water) from the underlying sediments that resulted in precipitation of framboidal pyrite mediated by bacterial processes at low temperature ($<50^{\circ}\text{C}$; **Machel, 2001; Hitzman et al., 2005; Dewaele et al., 2006**), within the carbonaceous sediments of the D'kar Formation (first mineralising stage). Dissolution of probable (as discussed above) evaporitic units within the Ngwako Pan Formation (red-beds) and possibly fluids developed due to compaction of evaporite-bearing D'Kar Formation (grey-beds) probably generated saline basinal fluids, as connate waters are reputed to be of high (10-30 wt.% equiv. NaCl) salinity (**Kesler et al., 1995; Kesler, 2005**). The dissolution of the evaporite also provided Cl^- , salinity, Mg^{2+} , Ca^{2+} , Na^{2+} , and SO_4^{2-} that gives the fluid the ability to dissolve and transport important amounts of copper (and possibly other trace metals) as chloride complexes under oxidising, near-neutral pH and low temperature conditions (**cf. Brown, 2005, 2009**) and also increases fluid density (**Selley et al., 2005**). These dense fluids with the help of basin faults and syn-sedimentary faults (see Fig. 4.6a) will tend to flow downwards through the Palaeoproterozoic granitoids and gneisses and the Mesoproterozoic Kgwebe Volcanics basement and the Ngwako Pan units and, as **Blundell et al. (2003)** proposed, probably seeped into a network of fractures in the basement where they attain elevated temperatures. This hot fluid ($\sim 400^{\circ}\text{C}$; as demonstrated in section 7.6.1 above) migrated through the sedimentary sequence leaching metal/copper and producing large-scale albitisation (see **Hitzman et al., 2005**). The migration of hot metalliferous/cupriferous brine into the overlying reduced beds (carbonaceous shales, siltstones, sandstones and carbonates) precipitated sulphides (second mineralising stage) across the redox boundary, with sulphur either supplied by biogenic pyrite (**Sun and Püttman, 1997; Hitzman et al., 2005**), reduced organic matter (**Sun and Püttman, 1997; Selley et al., 2005**), evaporites (**Ohmoto et al., 1990; Hitzman et al., 2005; Bull et al., 2011**), or hydrothermal fluids (**Hitzman et al., 2005**). As the basinal fluids endured fluid-rock interaction, the thermo- and physiochemical characteristics of the fluid were changing, resulting in a sequence of alteration patterns that include sodic, phyllic, propylitic, and carbonate stages. Mineralisation deposition occurs in a temperature range of 400 to $\leq 60^{\circ}\text{C}$ and at pH range of weakly acidic to near-neutral and alkaline conditions as constrained by carbonate or silicate equilibria (**cf. Robb, 2005**).

Subsequent closing of the rift during the Pan-African orogeny caused deformation and metamorphism throughout the region (**Key and Mapeo, 1999**). The resultant structures (e.g., faults, fractures and breccia zones) may have served as conduits for the syn-tectonic fluids that remobilised and redistributed the early formed sulphides into veins usually emplaced parallel to sub-parallel to the stratigraphy. Deformation of early formed sulphides initiated the growth of porphyroblasts and stretched and boudinaged sulphide mineralisation (third mineralising stage). It must be noted that the formation of bedding parallel veins requires extensional forces perpendicular to the bedding (**Brems et al., 2009**). With the progression of deformation (i.e., compression stage), these earlier emplaced veins were subsequently folded and crosscut by several new veins that are randomly oriented. The enrichment of sulphides in the hinge regions of the folds is evidence of the redistribution of ore-forming elements during subsequent metamorphism and deformation (cf. **Brems et al., 2009; Muchez et al., 2010**). Late stage massive veins that possibly formed during the latest phase of deformation crosscut all earlier formed veins. These veins occurred in dilation fractures in anticlinal fold hinges, which channelled and concentrated mineralisation.

During weathering, low-temperature meteoric fluids induced supergene processes re-precipitated metals in the supergene blanket (**Van Langendonck et al., 2013**), resulting in strong enrichment of chalcocite under the redox barrier and precipitation of carbonate hydroxides (malachite) above the redox barrier. This supergene process is further supported by the association of chalcocite with iron oxides/hydroxides indicating mineralisation temperature of below 103°C (**Ramdohr, 1980; Fleet, 2006**). These Cu-bearing phases are stable at surface conditions and in the range of pH 4 to 9 in the case of copper oxides (Anderson, 1982) (Fig. 7.8). A major factor that controls the formation of these minerals is the acidity of the environment (**Ridley, 2015**). Copper-rich phases such as native copper, chalcocite, digenite, and covellite can form at intermediate oxidation state in the presence of sulphur.

CHAPTER 8

GENERAL CONCLUSION AND PERSPECTIVES

This MSc thesis aims at constraining the petrogenesis of the Banana Zone Cu-Ag mineralisation and consequently understanding the processes that combined to form the mineralisation in the GCB, particularly in the Banana Zone area. The chief objectives of the study are; (1) to conduct petrological and geochemical assessment of the host rocks in order to classify the rocks, evaluate the weathering intensity at the source area and enhance the understanding of the sediments (siliciclastic) provenance and tectonic setting of the region, (2) to characterised the style of mineralisation in the Banana Zone, (3) to document ore and hydrothermal alteration mineral paragenesis in the Banana Zone in order to arrive at the processes that could have led to the mineralisation, and (4) eventually propose a comprehensive genetic model of the Banana Zone mineralisation that is likely to help in designing exploration models.

The results obtained and the discussions elaborated (see chapter 7) allow drawing the following conclusions

1. The Banana Zone Cu-Ag mineralisation is hosted in sediments from the Ngwako Pan and D'kar Formations. The sandstone units of both the Ngwako Pan and the D'Kar Formations are considered potassic and classified as sub-arkose and arkose.
2. The Banana Zone Cu-Ag mineralisation host rock sediments were largely supplied from a felsic igneous source rocks, namely the Kgwebe Volcanics, with subordinate sedimentary and metamorphic rocks and deposited in a continental rift setting (passive margin).
3. The Banana Zone Cu-Ag mineralisation is characterised by two main mineralisation styles that include dissemination and structure-controlled mineralisation and two modes of hydrothermal alteration occurrence including pervasive and structure-controlled modes. Alteration types include in order appearance: sodic-calcic, phyllic, silicification, propylitic and carbonatisation. Alteration patterns acts as controls on mineralisation. The Fe-S is predominantly

associated with phyllic varieties, whereas Cu-S mineralisation is mostly associated with Mg-chlorite rich zones (propylitic alteration) and carbonate-rich zones (carbonate alteration).

4. The main Cu-bearing minerals comprise of chalcopyrite, bornite, and chalcocite, along with minor or traces of digenite, covellite, and malachite. A few occurrences of sphalerite and galena are also observed. In addition to these, other phases include pyrite, arsenopyrite, pyrrhotite, cobaltite, tetrahedrite/tennantite, native bismuth, emplectite (?), and/or wittichenite (?).
5. Textural evidence and alteration patterns demonstrate a complex multistage history of sulphide precipitation spanning from syn-diagenetic through metamorphism to post-tectonic and supergene alteration. This protracted history of mineralisation involved multiple fluids of different origin and most probably of different composition that produced alteration of the rocks. The hypogene mineralising fluids were characterised by low to moderate temperature ($\leq 60^{\circ}\text{C}$ - 400°C), weakly acidic to near neutral and alkaline pH conditions. The fluids are considered to have been dense, oxidised, moderately saline (15-20 wt. % equiv. NaCl) and enriched in Na, K, Ca, and Mg and associated Cl ions. The supergene mineralising fluids were characterised by very low temperature ($\sim 25^{\circ}\text{C}$ - 103°C), oxidised, acidic groundwaters.
6. The fluids developed as a result of expulsion from deep in the sedimentary basin during burial diagenesis and migrated upward and towards the margins of the basin. The Ngwako Pan Formation (red bed) and the basement rocks of the Kgwebe Volcanics form the most viable metal source for the precipitation of sulphides in the area. The sulphur could have been derived from several sources that include biogenic pyrite, reduced organic matter, evaporites and hydrothermal fluids.

Although geochemical assays carried out by Khoemacau Mining ascertained the presence of Ag, no Ag-bearing mineral was observed in this study. Silver is possibly occurring in chalcocite as observed in the Dikulushi (Democratic Republic of Congo) Cu-Ag deposit (**Haest et al. 2007**) and the Polish Kupferschiefer deposit (**Large et al., 1995**). Furthermore, Ag is also likely to occur as isomorphic admixtures in tetrahedrite/tennantite and galena as observed in other hydrothermal systems (**Bineli Betsi et al., 2013**). The

recommendation is to carry out a Laser ablation-inductively coupled plasma spectrometry (LA-ICP-MS) analysis in order to map out silver occurrence in these minerals and therefore unravel the mode of occurrences of Ag mineralisation not only within the Banana Zone, but also throughout the Khoemacau Mining Ghanzi Project.

Furthermore, in this study, alteration mineral assemblages were used to estimate the temperature of mineralising fluids. As a consequence, this estimate was based on the stability field of mineral assemblages from porphyry copper deposits, which form at higher temperature compared to the sediment-hosted copper deposits. To better constrain the temperature of the fluids it is recommended that stable isotope (S, O) studies be conducted to determine the temperature. In addition to this, sphalerite and chlorite geothermometry should be performed. Also, the age of the mineralisation is poorly constrained, as the previous geochronology data by **Hall (2013)** at Boseto Mine was based on Re-Os model age. The disadvantage of using model age to constrain the age is that in this model, it is assumed that there is no common osmium and all the osmium present is radiogenic. This could give incorrect ages. Therefore, to better constrain the age of mineralisation conventional dating methods, such as $^{40}\text{Ar}/^{39}\text{Ar}$ are recommended for dating alteration minerals, such as adularia and illite to obtain the radiometric dates on the wall rock alteration and, by inference, the age of mineralisation. Lastly, but not least, alteration pattern distribution maps are recommended to show distinct areas characterised by particular alteration patterns in order to assess the hydrothermal fluid evolution at property scale.

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APPENDIX A1

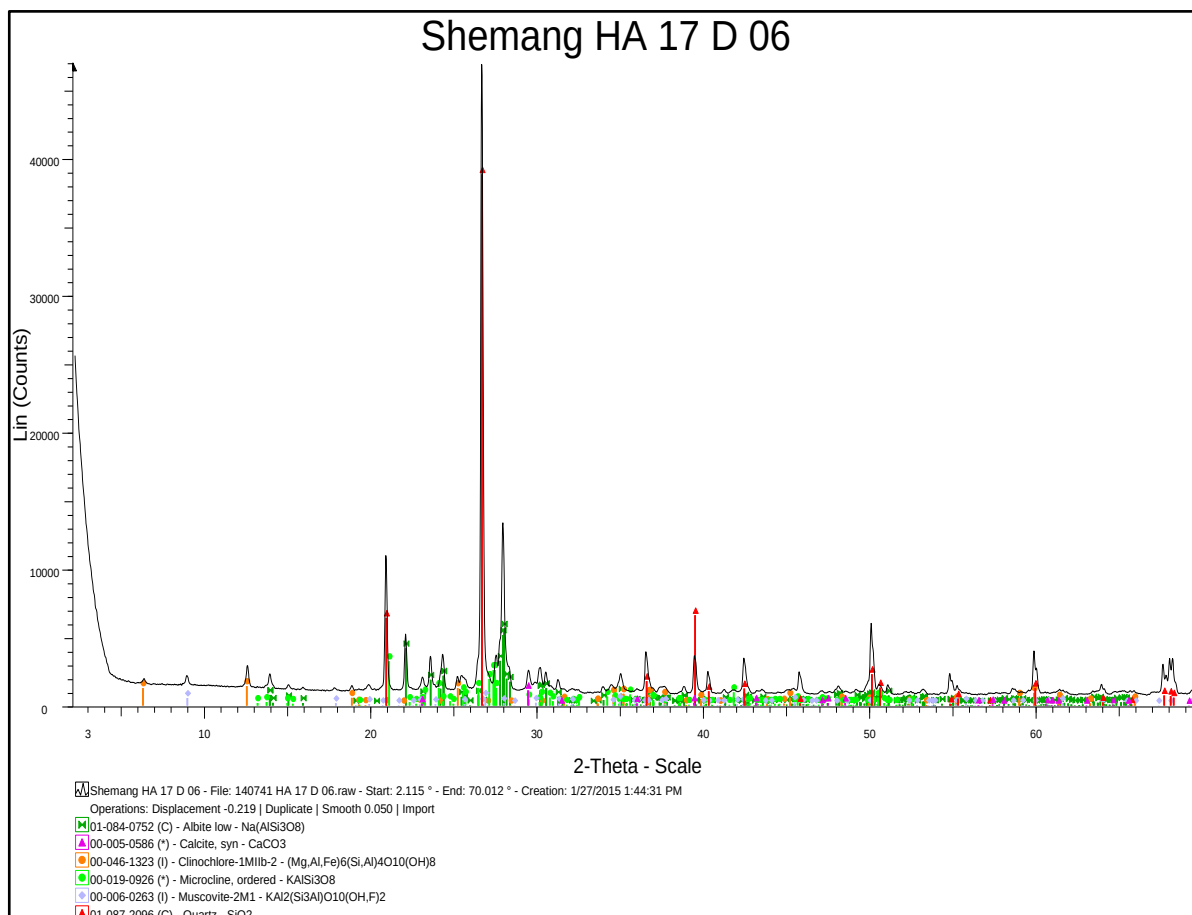
DRILL HOLE COLLAR LOCATIONS

Table A1 Exploration drill hole collar locations

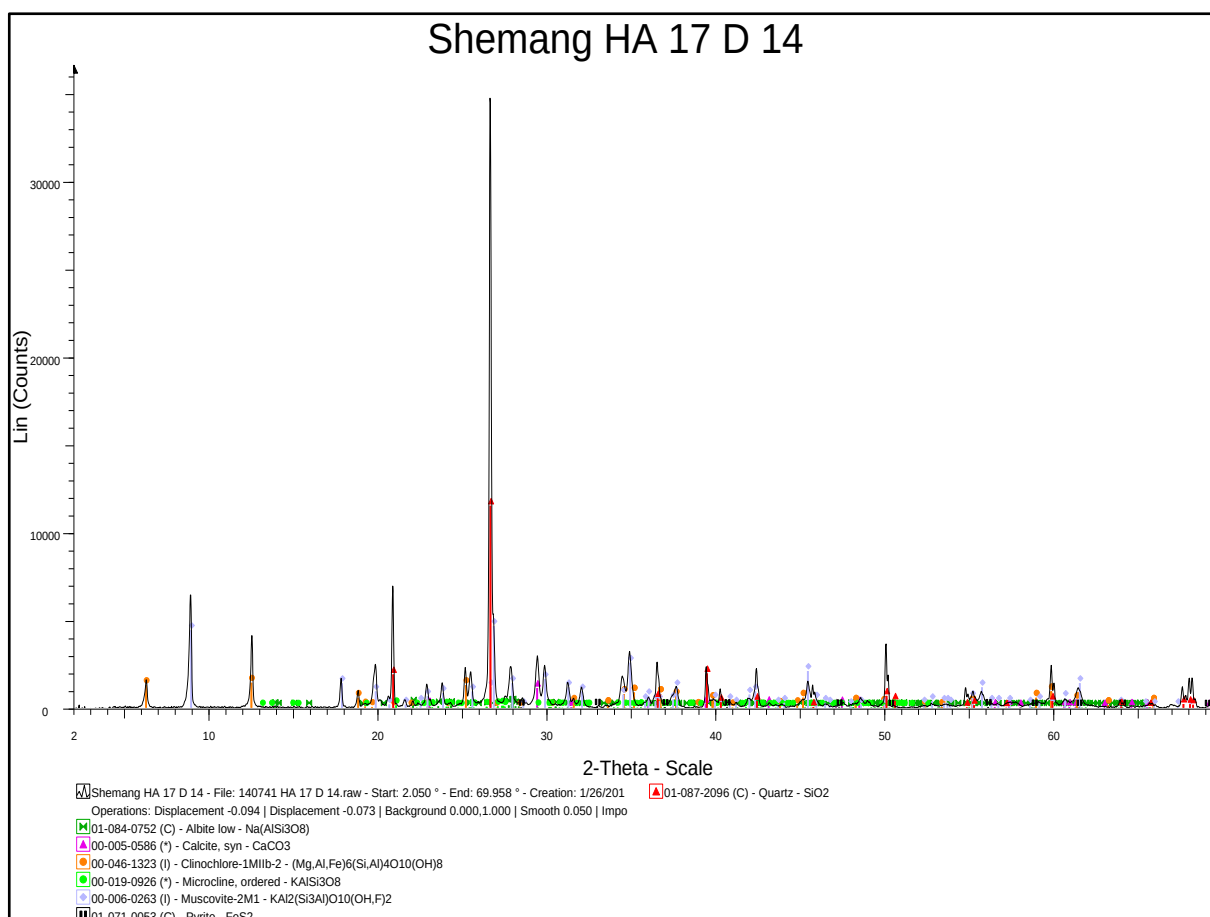
ZONE	DRILL HOLE I.D.	EASTING	NORTHING
Chalcocite	HA 515-D	666372.60	7658880.00
	HA 520-D	664753.40	7658726.00
	HA 524-D	663966.50	7660146.00
New Discovery	HA 244-D	673635.00	7666281.09
	HA 248-D	672844.50	7666031.00
	HA 251-D	673241.80	7666516.00
	HA 500-D	672925.90	7665915.70
North Limb	HA 17-D	687791.08	7675877.11
	HA 38-D	683385.32	7672800.99
	HA 90-D	686284.73	7674696.27
North East Fold	HA 10-D	689030.01	7676533.74
	HA 487-D	689913.54	7676861.26
	HA 526-D	689712.90	7676881.00
South Limb Definition	HA 430-D	686152.59	7672042.48
	HA 481-D	685087.54	7670999.98
	HA 486-D	684852.83	7670814.28

APPENDIX A2

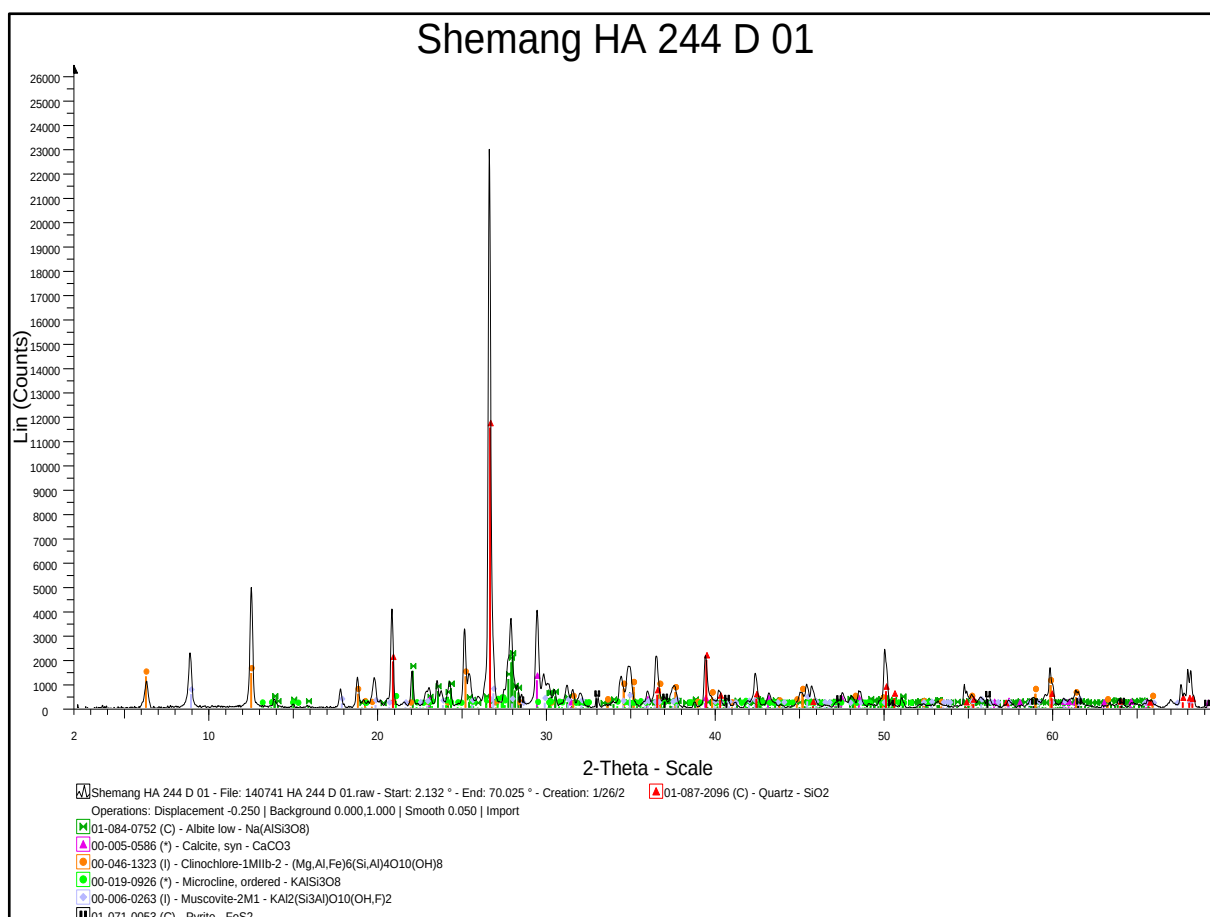
X-RAY DIFFRACTOMETRY DIFFRACTOGRAMS FOR THE SELECTED SAMPLES



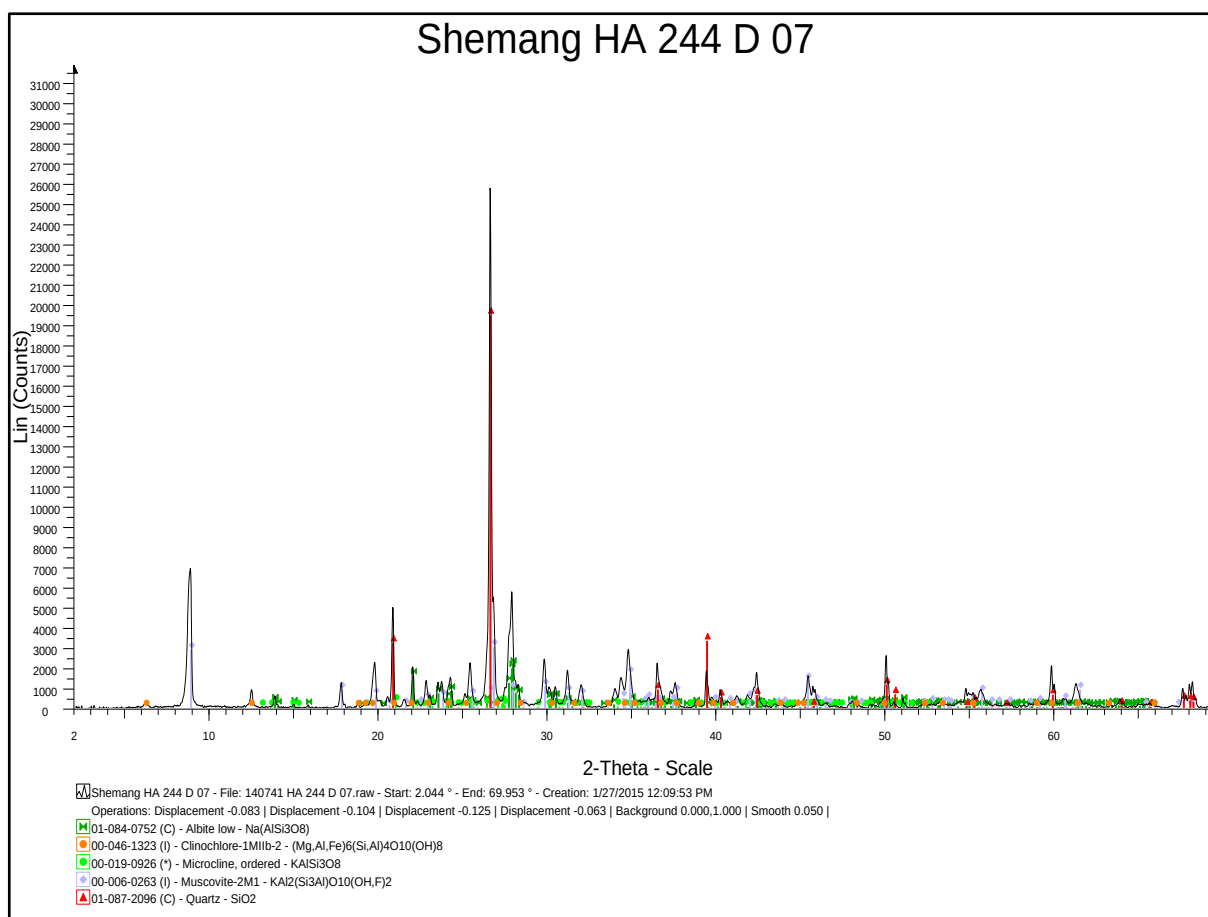
Samples scanned from 2 to 70° 2θ Cu_{Kα} radiation □□ at a speed of 0.02° 2θ □ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



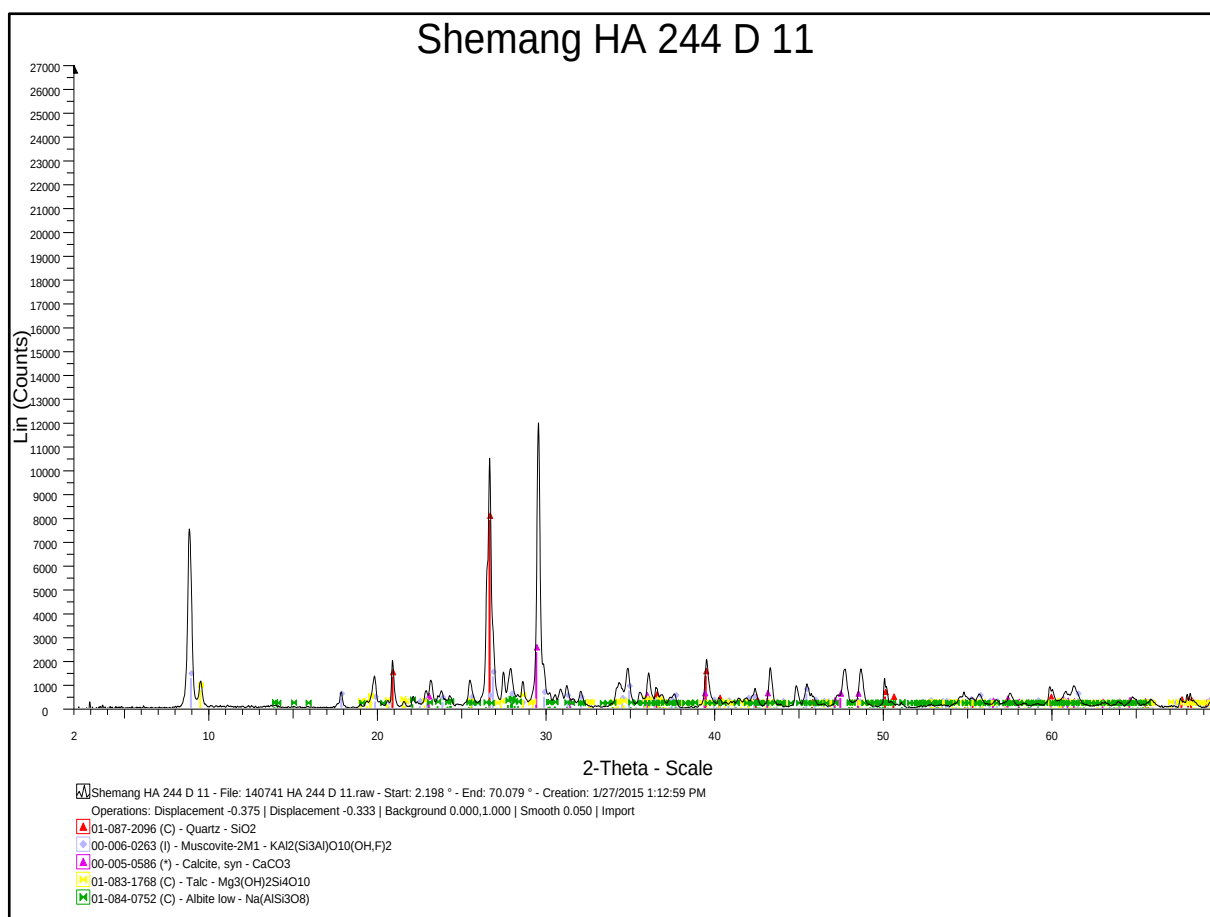
Samples scanned from 2 to 70° 2θ Cu_{Kα} radiation □□ at a speed of 0.02° 2θ □ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



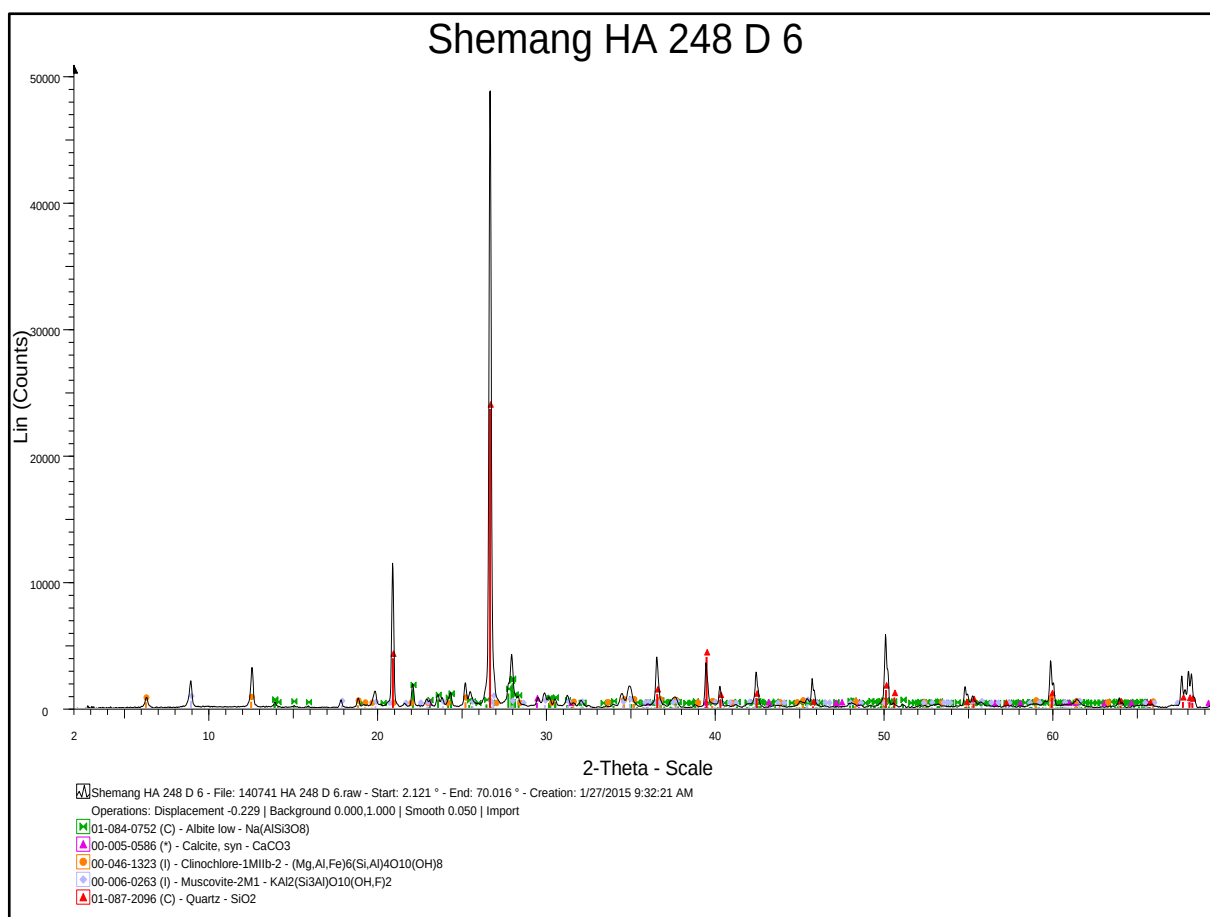
Samples scanned from 2 to 70° 2θ CuK_α radiation at a speed of 0.02° 2θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



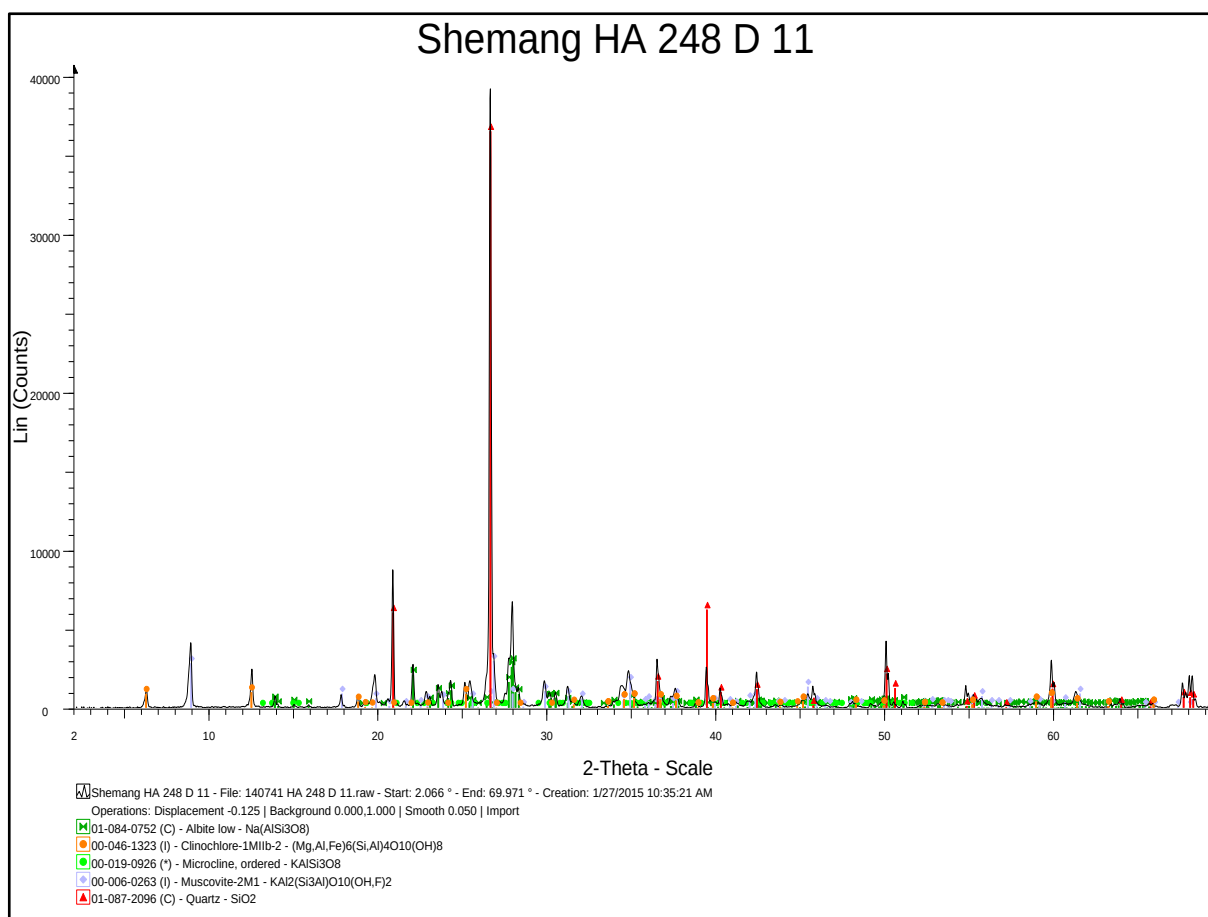
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



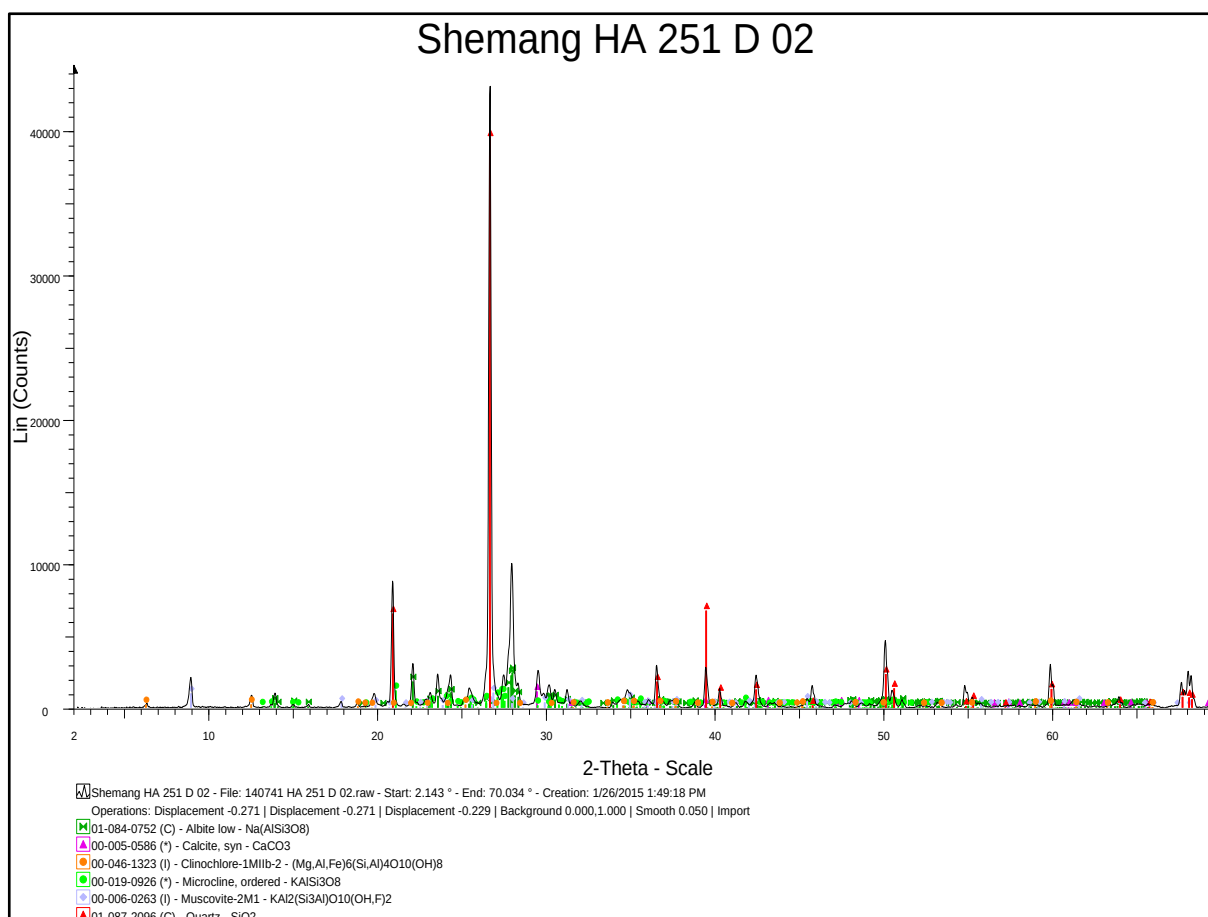
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



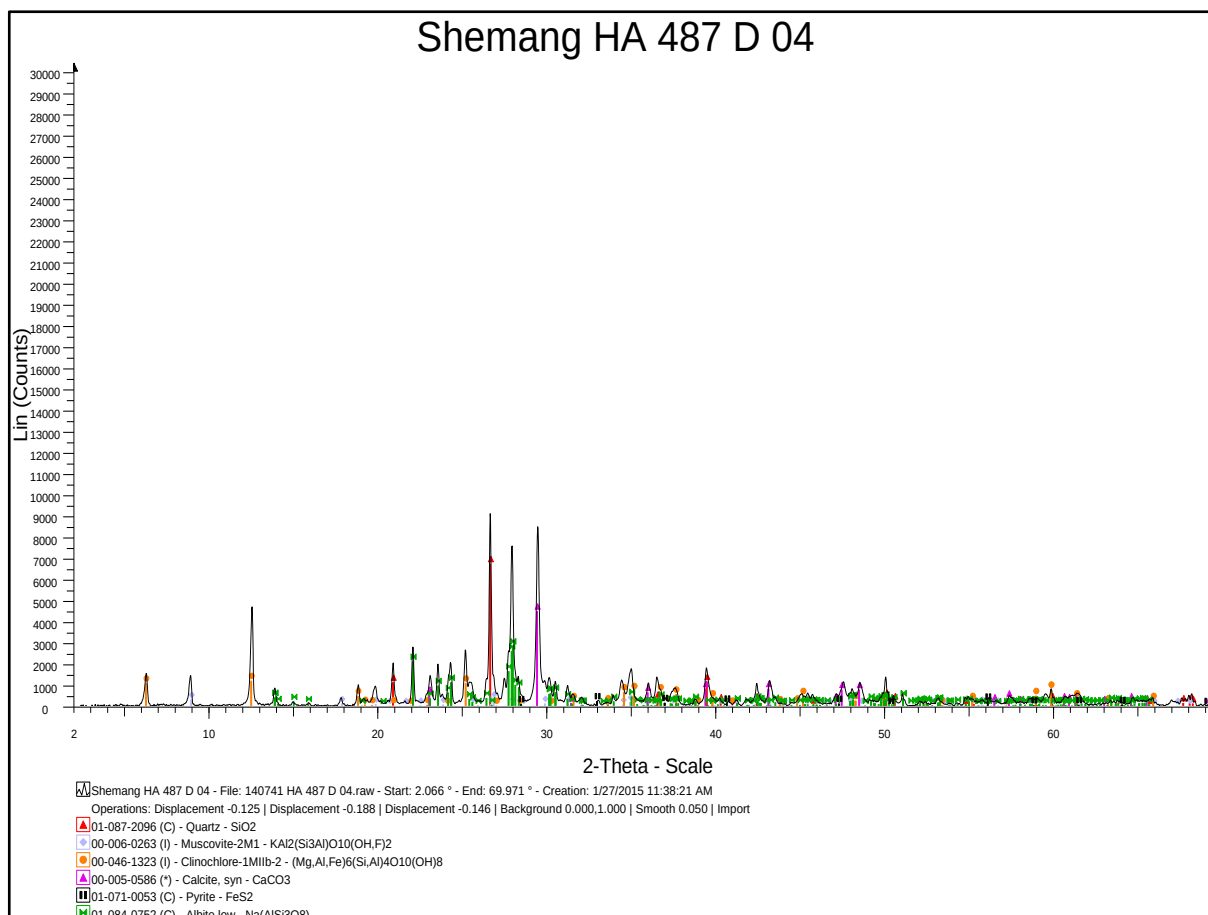
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



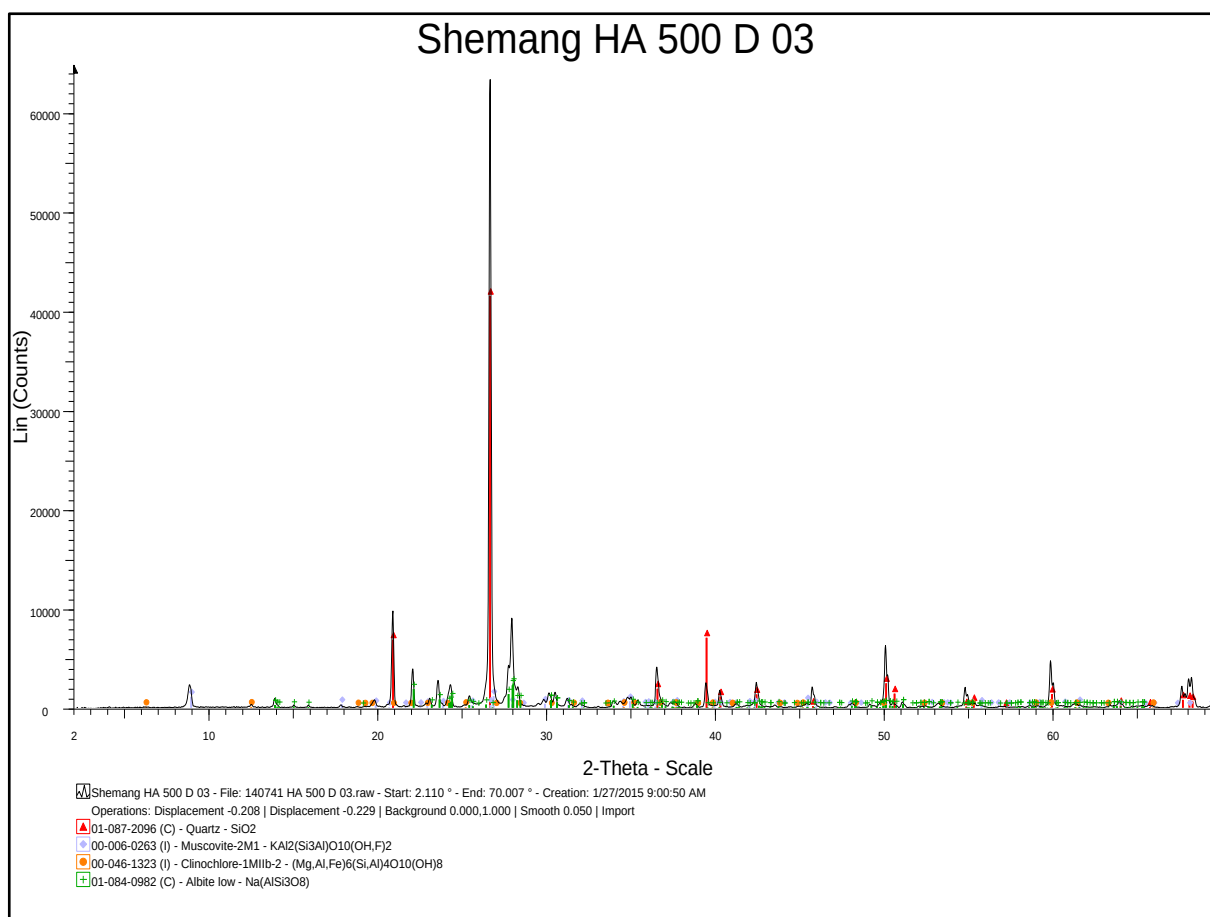
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



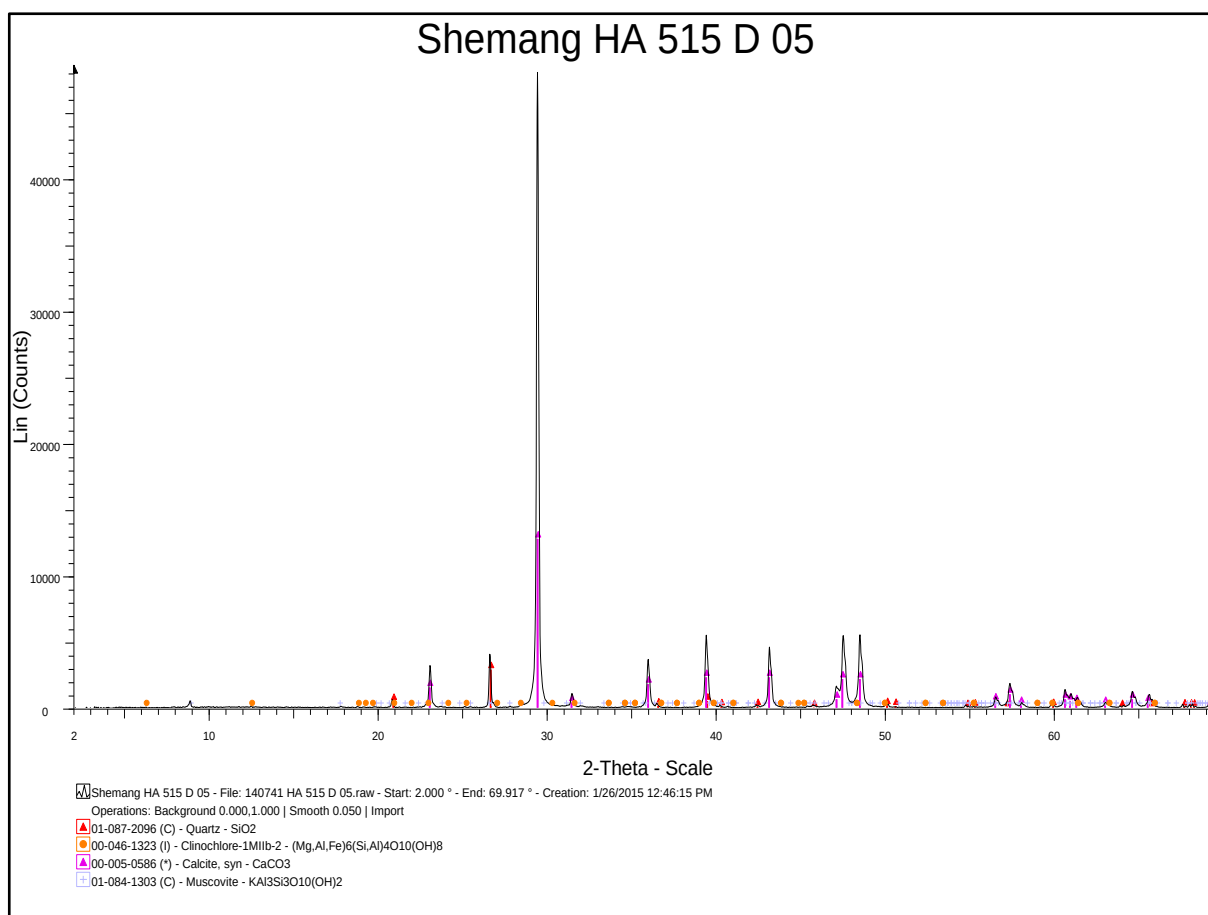
Samples scanned from 2 to 70° 2θ Cu_{Kα} radiation □□ at a speed of 0.02° 2θ □ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



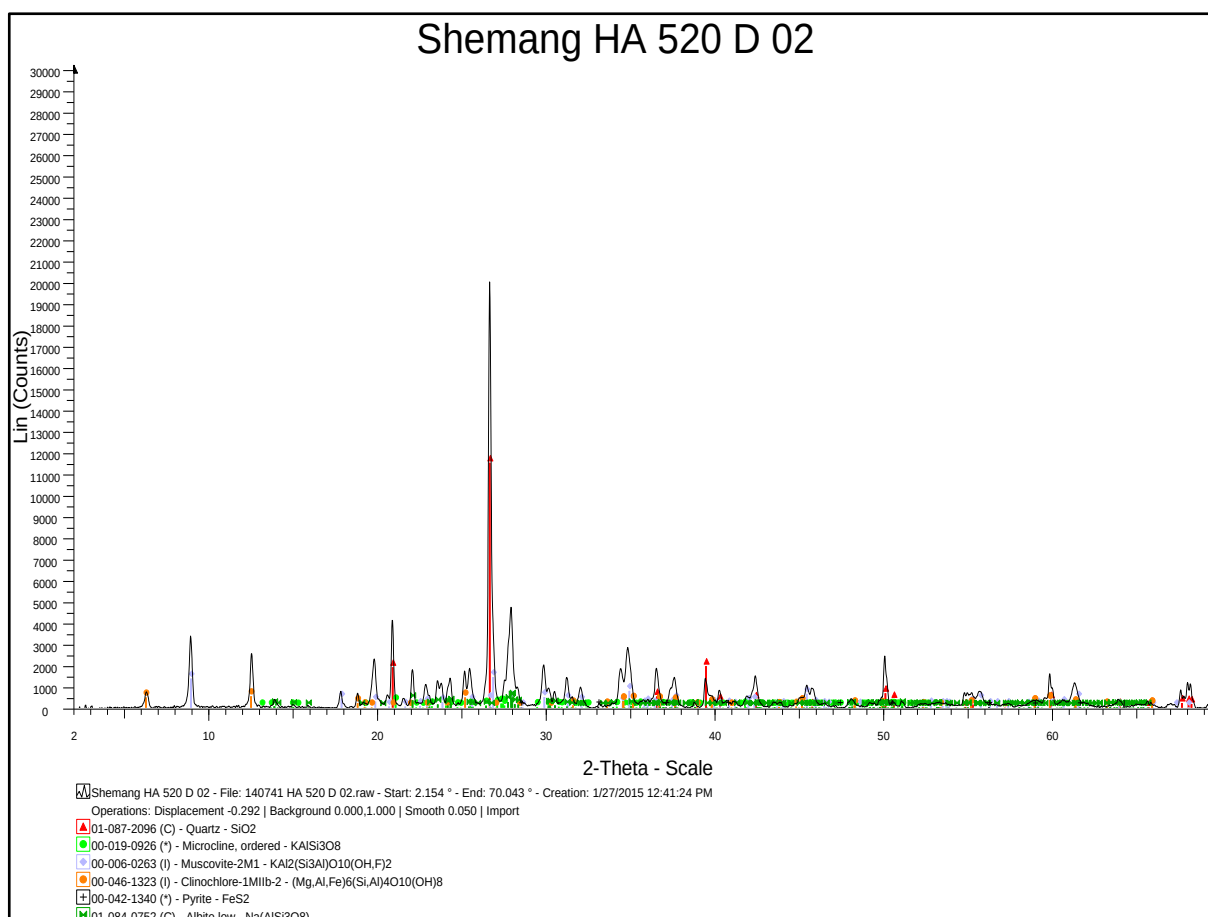
Samples scanned from 2 to 70° 2 θ $\text{CuK}\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



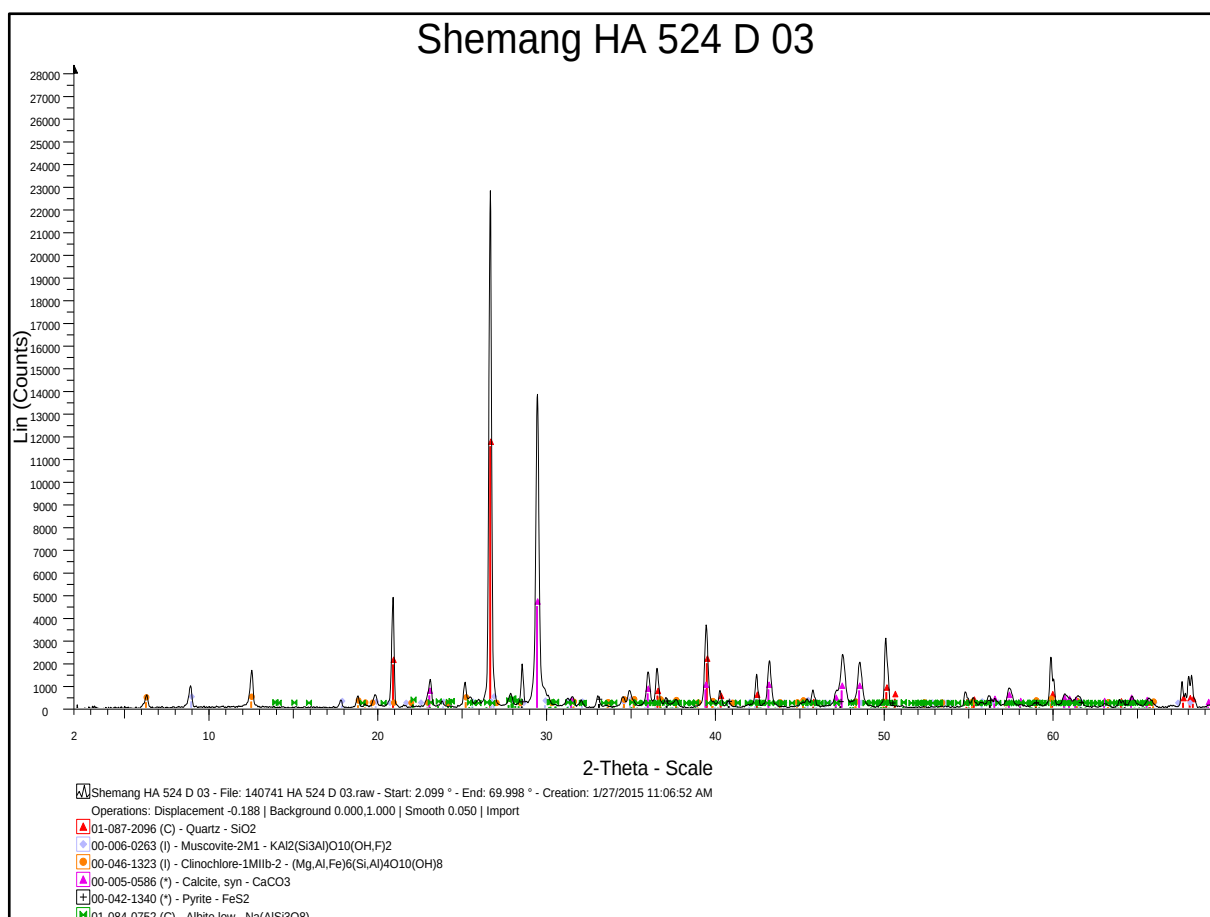
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



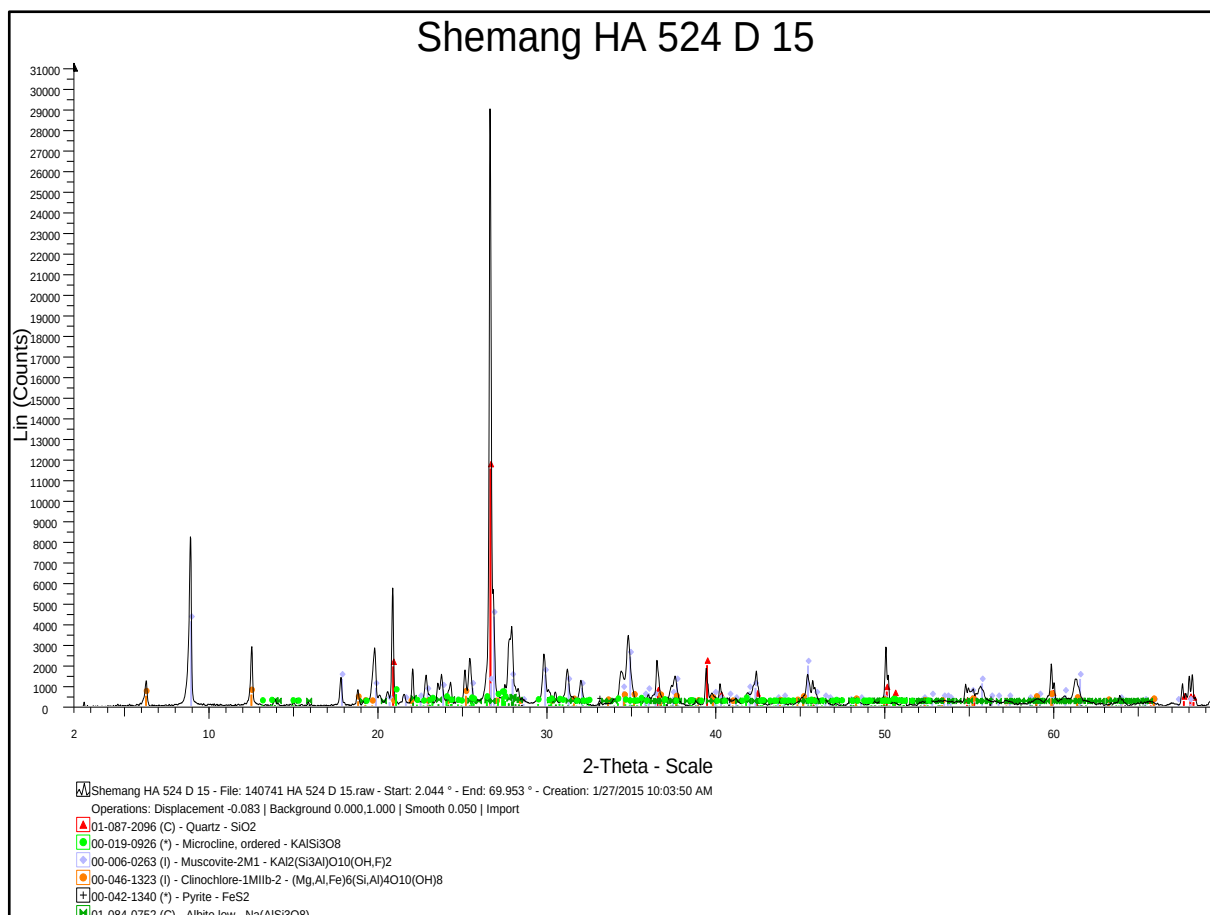
Samples scanned from 2 to 70° 2 θ Cu K_{α} radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



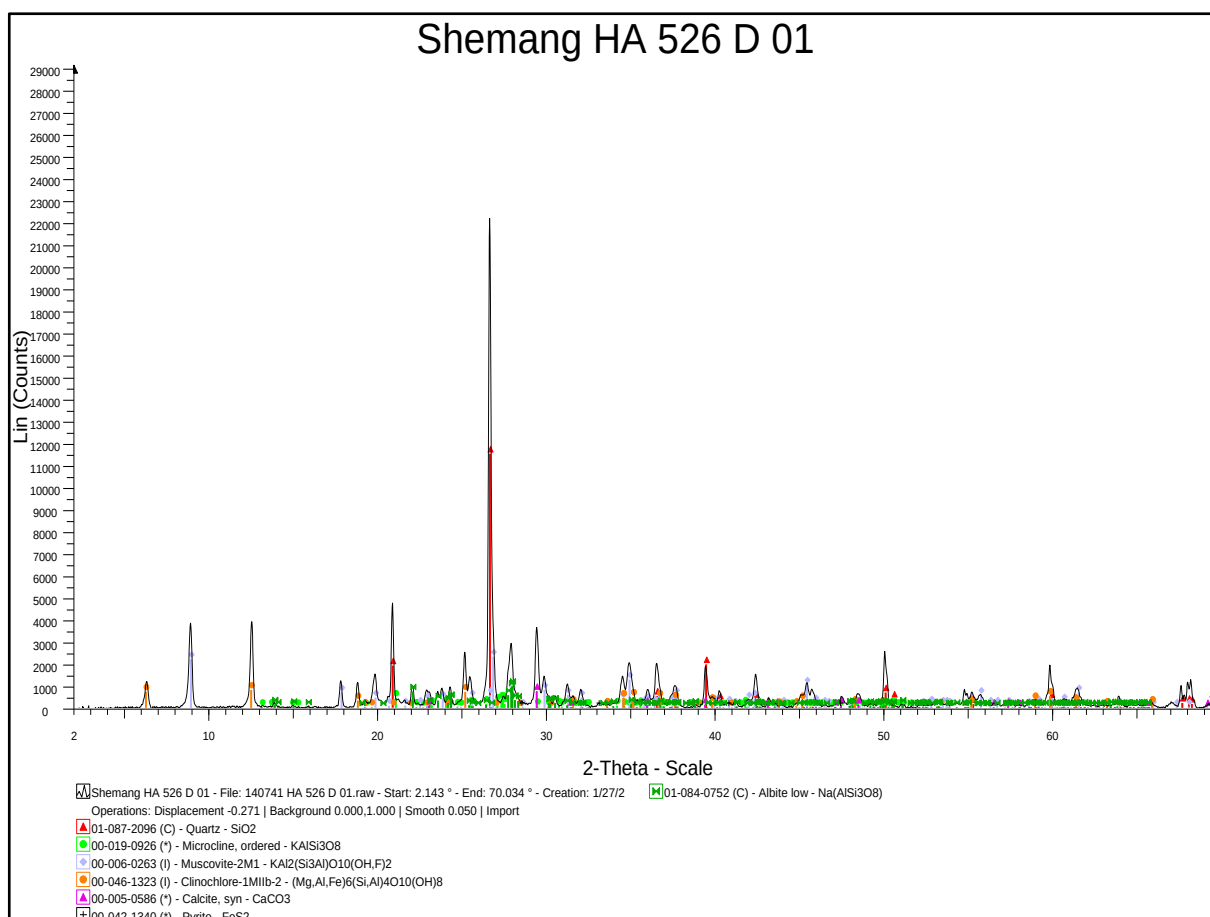
Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



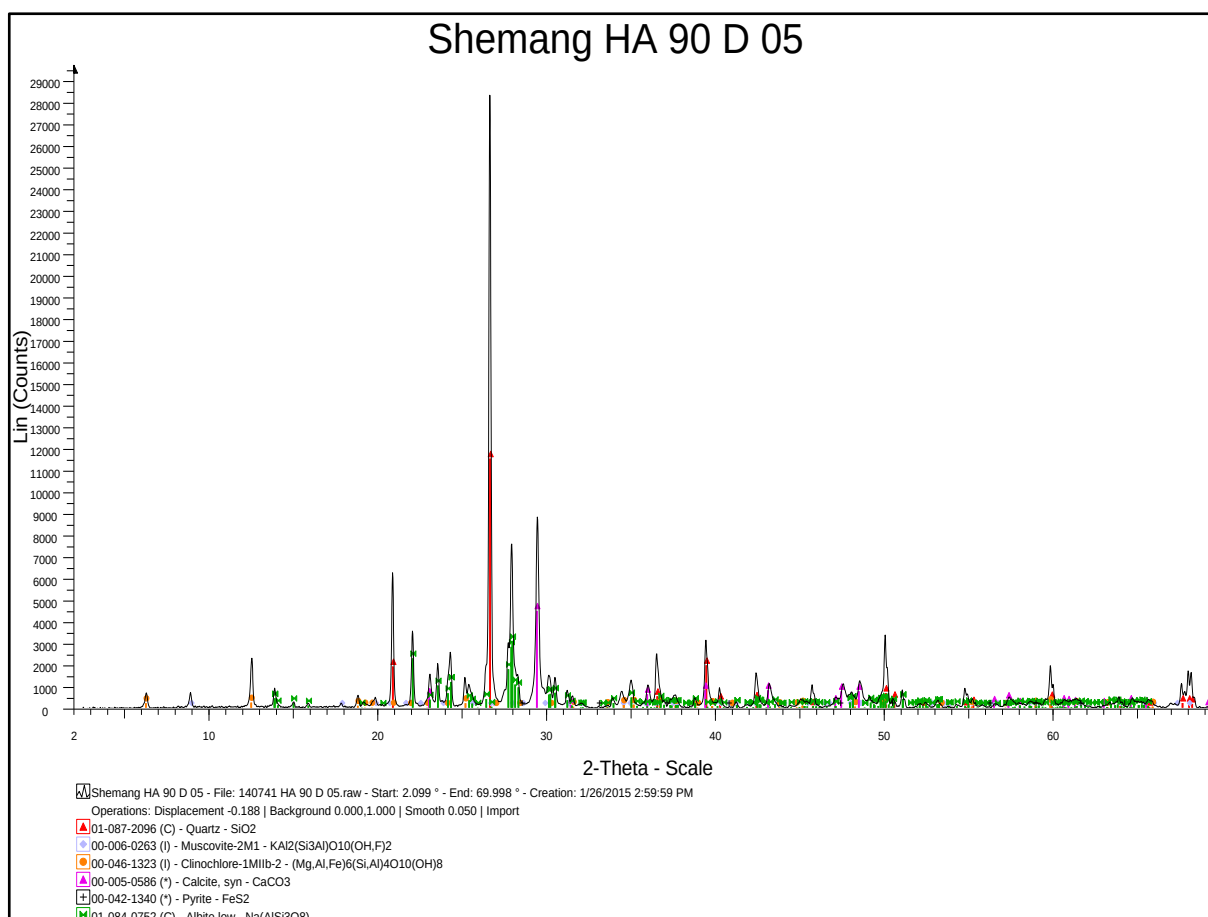
Samples scanned from 2 to 70° 2 θ Cu K_{α} radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



Samples scanned from 2 to 70° 2 θ $\text{CuK}\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).



Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brine, 1985).

APPENDIX A3

PHASE CONCENTRATIONS

Sample	Calcite	Pyrite	Talc	Microcline	Plagioclase	Quartz	Chlorite	Mica
HA-17-D-06	3	-	-	4	28	49	10	5
HA-17-D-14	6	tc	-	tc	2	31	24	35
HA-244-D-01	10	2	-	tc	10	26	34	16
HA-244-D-07	-	-	-	2	14	30	6	47
HA-244-D-11	26	-	8	-	3	16	-	47
HA-248-D-11	-	-	-	2	17	40	15	27
HA-248-D-06	tc	-	-	-	13	51	21	14
HA-251-D-02	5	-	-	5	21	49	6	14
HA-487-D-04	21	tc	-	-	21	11	35	11
HA-500-D-03	-	-	-	-	22	59	2	17
HA-515-D-05	92	-	-	-	-	4	tc	3
HA-520-D-02	-	tc	-	3	15	28	23	30
HA-524-D-03	40	2	-	-	2	33	14	9
HA-524-D-15	-	tc	-	2	11	26	16	45
HA-526-D-01	8	tc	-	tc	7	27	28	28
HA-90-D-05	20	tc	-	-	25	33	16	6

Note; tc is trace amount

Samples scanned from 2 to 70° 2 θ Cu $K\alpha$ radiation at a speed of 0.02° 2 θ steps size/0.5 sec, LYNXEYE detector and generator settings of 40 kV and 40mA. Phase concentrations are determined as semiquantitative estimates, using relative peak heights/areas proportions (Brime, 1985).

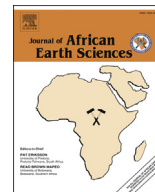
APPENDIX A4

Table A4 Adjusted major and minor elements composition (on an anhydrous basis) and their isometric log-transformed ratios for the Ngwako Pan Formation sandstone, D’Kar Formation sandstone and D’Kar Formation mudrocks

	Ngwako Pan Formation sandstone						D’Kar Formation sandstone						D’Kar Formation mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg	MS1	MS2	MS3	MS4	MS5	Avg	MD1	MD2	MD3	MD4	MD5	Avg
Major and minor elements (wt. %)																		
(SiO₂)_{adj}	77.54	77.78	79.62	78.29	78.95	78.43	77.59	77.38	77.34	76.25	77.60	77.23	66.04	66.70	65.64	64.33	62.57	65.06
(TiO₂)_{adj}	0.21	0.40	0.30	0.31	0.30	0.30	0.29	0.47	0.36	0.42	0.46	0.40	0.82	0.78	0.69	0.76	0.89	0.79
(Al₂O₃)_{adj}	7.84	9.85	8.70	8.81	8.75	8.79	8.52	10.18	9.81	9.65	10.02	9.64	16.42	15.06	12.83	16.18	18.28	15.75
(Fe₂O₃ (t))_{adj}	0.96	1.98	1.58	1.51	1.55	1.52	1.12	2.44	1.71	2.17	2.33	1.95	5.27	5.77	5.82	6.90	5.52	5.86
(MnO)_{adj}	0.17	0.09	0.09	0.11	0.10	0.11	0.12	0.08	0.09	0.10	0.09	0.10	0.11	0.15	0.26	0.12	0.11	0.15
(MgO)_{adj}	0.37	0.74	0.42	0.52	0.46	0.50	0.58	0.71	0.61	0.52	0.62	0.61	2.92	2.74	4.11	3.08	3.45	3.26
(CaO)_{adj}	7.36	2.93	3.78	4.69	4.23	4.60	6.15	2.57	3.92	3.66	3.16	3.89	1.59	2.56	5.71	2.58	1.51	2.79
(Na₂O)_{adj}	1.86	2.65	2.36	2.30	2.33	2.30	2.13	3.05	2.45	2.27	2.69	2.52	2.03	2.91	3.02	1.98	2.94	2.57
(K₂O)_{adj}	3.62	3.46	3.07	3.39	3.23	3.35	3.42	2.99	3.62	2.86	2.95	3.17	4.60	3.14	1.73	3.86	4.50	3.57
(P₂O₅)_{adj}	0.06	0.09	0.07	0.07	0.07	0.07	0.07	0.11	0.09	0.07	0.09	0.09	0.20	0.19	0.18	0.19	0.22	0.19
Isometric log-transformed ratios																		
ln(TiO₂/SiO₂)_{adj}	-5.90	-5.26	-5.58	-5.53	-5.58	-5.57	-5.59	-5.10	-5.36	-5.20	-5.13	-5.27	-4.39	-4.45	-4.55	-4.44	-4.25	-4.42
ln(Al₂O₃/SiO₂)_{adj}	-2.29	-2.07	-2.21	-2.18	-2.20	-2.19	-2.21	-2.03	-2.06	-2.07	-2.05	-2.08	-1.39	-1.49	-1.63	-1.38	-1.23	-1.42
ln(Fe₂O₃/SiO₂)_{adj}	-4.39	-3.67	-3.92	-3.95	-3.93	-3.97	-4.24	-3.46	-3.81	-3.56	-3.51	-3.72	-2.53	-2.45	-2.42	-2.23	-2.43	-2.41
ln(MnO/SiO₂)_{adj}	-6.13	-6.73	-6.75	-6.54	-6.64	-6.56	-6.46	-6.85	-6.72	-6.63	-6.73	-6.68	-6.36	-6.13	-5.55	-6.24	-6.31	-6.12
ln(MgO/SiO₂)_{adj}	-5.34	-4.65	-5.24	-5.02	-5.14	-5.08	-4.89	-4.69	-4.84	-4.98	-4.84	-4.85	-3.12	-3.19	-2.77	-3.04	-2.90	-3.00
ln(CaO/SiO₂)_{adj}	-2.35	-3.28	-3.05	-2.82	-2.93	-2.88	-2.54	-3.40	-2.98	-3.04	-3.20	-3.03	-3.73	-3.26	-2.44	-3.22	-3.72	-3.27
ln(Na₂O/SiO₂)_{adj}	-3.73	-3.38	-3.52	-3.53	-3.52	-3.54	-3.59	-3.23	-3.45	-3.51	-3.36	-3.43	-3.48	-3.13	-3.08	-3.48	-3.06	-3.25
ln(K₂O/SiO₂)_{adj}	-3.07	-3.11	-3.25	-3.14	-3.20	-3.15	-3.12	-3.25	-3.06	-3.28	-3.27	-3.20	-2.66	-3.06	-3.63	-2.81	-2.63	-2.96
ln(P₂O₅/SiO₂)_{adj}	-7.11	-6.73	-7.01	-6.99	-7.00	-6.97	-6.97	-6.53	-6.72	-6.99	-6.73	-6.79	-5.82	-5.88	-5.89	-5.84	-5.66	-5.82

APPENDIX A5

This appendix presents Tebogo Kelepile, Thierry Bineli Betsi, Fulvio Franchi, Elisha Shemang, Cheo Emmanuel Suh (2017) in its published format. As a result, the formatting, layout, figures and tables numbering do not follow the layout of this thesis.



Provenance and tectonic setting of the Neoproterozoic clastic rocks hosting the Banana Zone Cu-Ag mineralisation, northwest Botswana



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ABSTRACT

Petrographic and geochemical data were combined in order to decipher the petrogenesis of the Neoproterozoic sedimentary succession associated with the Banana Zone Cu-Ag mineralisation (northwest Botswana), in the Kalahari Copperbelt. The investigated Neoproterozoic sedimentary succession is composed of two formations including the Ngwako Pan and the D'kar Formations. The Ngwako Pan Formation is made up of continental siliciclastic sediments, mainly sandstones interbedded with siltstones and mudstones, whereas the D'kar Formation is comprised of shallow marine laminated siltstones, sandstones and mudstones, with subordinate limestone. Copper-Ag mineralisation is essentially confined at the base of the D'kar Formation, which bears reduced organic components, likely to have controlled Cu-Ag precipitation.

Sandstones of both the Ngwako Pan and the D'kar Formations are arkoses and subarkoses, composed of quartz (Q), feldspars (F) and lithic fragments (L). Moreover, geochemically the sandstones are considered as potassic and classified as arkoses. On the other hand, mudrocks of the D'kar Formation are finely laminated and are dominated by muscovite, sericite, chlorite and quartz.

The modified chemical index of weathering (CIW') values indicated an intense chemical weathering of the source rock. The dominance of detrital quartz and feldspar grains coupled with Al_2O_3/TiO_2 ratios (average 29.67 and 24.52 for Ngwako Pan and D'kar Formations, respectively) and Ni and Cr depletion in the sandstones, suggest a dominant felsic source. However, high concentrations of Ni and Cr and a low Al_2O_3/TiO_2 ratio (<20) in the mudrocks of the D'kar Formation indicate a mixed source. Provenance of the investigated sandstones and mudrocks samples is further supported by the REE patterns, the size of Eu anomaly as well as La/Co, Th/Co, Th/Cr and Cr/Th ratios, which show a felsic source for the sandstones of both the Ngwako Pan and D'kar Formations and an intermediate source for the mudrocks of the D'kar Formation. Detrital modes (QFL diagrams) and geochemical characteristics of the sandstones of both the Ngwako Pan and D'kar Formations indicate that the detritus were probably supplied from a heavily weathered felsic continental block and deposited in a continental rift setting (passive margin) in a humid environment. The source rocks might have been the Palaeoproterozoic basement rocks (granitoids and granitic gneiss) and the Mesoproterozoic Kgwebe volcanic rocks exposed north of the study area.

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1. Introduction

Combined petrographic and geochemical data of clastic sediments have been proved to be a powerful tool in deciphering the provenance and tectonic setting of sedimentary basin (Dickinson

and Suczek, 1979; Bhatia and Crook, 1986; McLennan et al., 1993; Cingolani et al., 2003; Das et al., 2008). The overall composition of siliciclastic sediments is influenced by several factors that include the source area composition, climate, sediment transport processes, burial, and diagenesis (e.g., Johnsson, 1993; Armstrong-Altrin et al., 2004, 2012). In petrographic studies, the modal compositions of sandstones and the nature of quartz grains as well as lithic fragments can be utilised to infer the type of source rocks

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(Folk, 1974, 1980; Blatt et al., 1980; Pettijohn et al., 1987; Asiedu et al., 2000) and their tectonic setting (Dickinson and Suczek, 1979; Dickinson et al., 1983; Weltje, 2002; Basu, 2003). However, it must be noted that grain size variations due to weathering processes can alter the mineralogical composition of sandstones and this can result in misinterpretation of provenance studies (Ingersoll et al., 1984 and reference therein). Similarly, some geochemical ratios can be altered during weathering (Taylor and McLennan, 1985) and/or diagenesis (Nesbitt and Young, 1989; Milodowski and Zalasiewicz, 1991), but as long as the bulk composition of a rock is not totally altered, geochemical analysis is a valuable tool in the study of sandstones (McLennan et al., 1993). Therefore, major oxides ratios (Roser and Korsch, 1986; Verma and Armstrong-Altrin, 2013, 2016) trace (Cr, Co, Ni, Th and Zr) and rare earth elements (REE) (Taylor and McLennan, 1985; McLennan, 2001, 2003; Wang and Zhou, 2012) distribution and their ratios (La/Co, Th/Co, Th/Cr, and Cr/Th) (Cullers et al., 1988; Wronkiewicz and Condie, 1989; Condie and Wronkiewicz, 1990; Cullers, 1994; McLennan, 2001; McLennan et al., 1990, 1993; Armstrong-Altrin et al., 2004,

2013) can be utilised for this purpose.

The Banana Zone Cu-Ag deposit (hereafter referred to as Banana Zone, Fig. 1), is part of the SW-NE trending Kalahari Copperbelt (Fig. 1a) that forms the southwestern annex of the Central African Copperbelt (CAC) in Botswana (Ghanzi-Chobe Belt) and in Namibia (Damara Belt). The Banana Zone (by analogy to its banana-like shape, see Fig. 1b) is an exploration site in the Ghanzi Project area owned by Khoemaqu Copper Mining and consists in a simple doubly plunging folded structure with no major faulting (GeoLogix, 2009). The Ghanzi Project is composed of three main mineralised zones, including from south to north the Banana Zone, Zone 5 and Boseto Mine (Fig. 1b). The Banana Zone, which is the focus of this study has inferred resource worth over 225.4 million metric tons (Mt) grading 0.64% Cu and 8.1 g/t Ag (Hana Mining Limited, 2011). The Banana Zone is within one of the world's most extensive metallogenic provinces of sediment-hosted copper deposits (Sillitoe et al., 2010) and despite the economic importance of this area, a thorough description of the petrology and geochemistry of the mineralised host rocks is

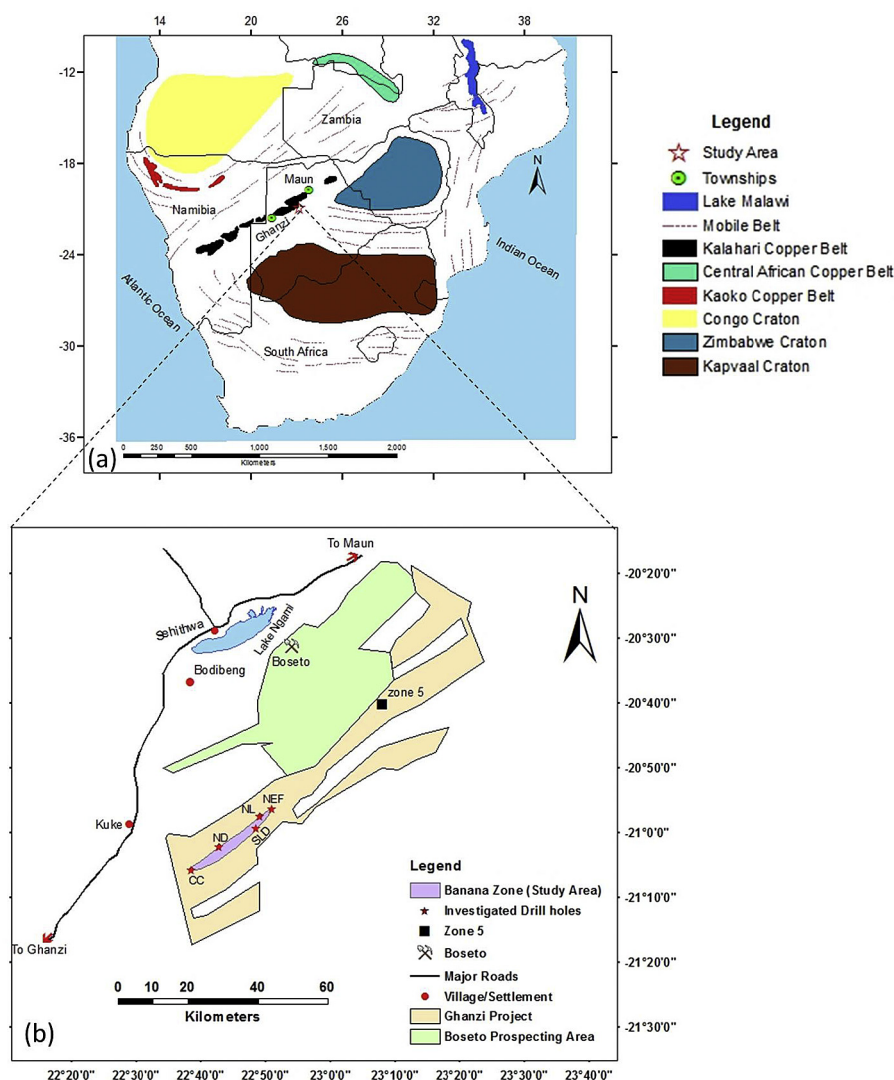


Fig. 1. (a). Schematic geological and structural map of southern Africa showing the location of the Ghanzi-Chobe Belt and the Banana Zone in relation to the Congo, Zimbabwe and Kaapvaal Cratons (modified from Maiden and Borg, 2011). (b) Outline of the Banana Zone (red star in (a)) showing the investigated drill holes location and the different mineralised sites: CC=Chalcocite Zone, ND=New Discovery, NL=North Limb, NEF=North East Fold, SLD=South Limb Definition. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

lacking and the provenance and tectonic setting of the sedimentary succession that hosts the Banana Cu-Ag deposit is not well understood. Previous works conducted in the Kalahari Copperbelt, mainly focused on the lithostratigraphy, depositional environment, and the description of the mineralised horizon in a broad sense (Borg, 1988; Borg and Maiden, 1989; Modie, 1996, 2000; Sillitoe et al., 2010).

In this paper, an integrated study of the petrography and geochemistry of the sediments pertaining to the Ghanzi Group (Fig. 2) is presented. This study aims at highlighting the characteristics of host rocks of the Banana Zone Cu-Ag deposit, their tectonic setting and environmental conditions at the time of deposition. Consequently, the objective of this study is to utilise petrographic and geochemical data in order to: (a) determine the source rock characteristics, (b) evaluate the weathering intensity at the source area and (c) infer the provenance and tectonic setting of the source area. This is of significant importance considering that the chemical composition, the mineralogical composition and the redox state may control metal precipitation and the patterns of hydrothermal alteration and therefore provide better information to understand the provenance and as well as the processes that converge to form the Cu-Ag mineralisation.

2. Geological setting

The Banana Zone Cu-Ag deposit is within the Ghanzi-Chobe Belt, which extends from the Namibia-Botswana border in the south, to Shinamba Hills in the north, near the Botswana-Zambia

border, trending in NE-SW direction (Kampunzu et al., 1998; Modie, 2000). The Ghanzi-Chobe Belt is composed of three lithostratigraphic units that include the Ghanzi Group and two informal units named Raibok Complex and Xaudum-Tsodilo Hills-Kaonaka Group (Key and Ayres, 1998; Kampunzu et al., 2000). In Botswana the Late Mesoproterozoic to Neoproterozoic Ghanzi-Chobe Belt is buried by the Phanerozoic cover, except in the Ghanzi Ridge and in the Goha and Shinamba Hills (Fig. 1) (Modie, 1996). The rocks of this belt were deposited in a volcano-sedimentary basin along the northern margin of the Kalahari Craton (Borg and Maiden, 1987; Modie, 1996). The basement rocks in this region are represented by: (i) Palaeoproterozoic granites, granitic gneisses and rhyolites (Okwa valley and Xangwa area), (ii) Mesoproterozoic Kgwebe volcanic rocks consisting of meta-rhyolites and minor metabasalts, and (iii) Pre-Neoproterozoic granitoids and high grade metamorphic rocks (Carney et al., 1994; Kampunzu et al., 2000).

The Banana Zone Cu-Ag deposit is hosted by the Neoproterozoic succession of the Ghanzi Group (Fig. 2). These sediments have been deposited in an intracontinental rift basin that formed as a consequence of the break-up of the Rodinia Supercontinent (Modie, 1996). The Ghanzi Group is divided into four formations including, from the base to the top: (i) the Kuke Formation, mainly composed of quartz arenite and red sandstone with a basal conglomerate; (ii) the Ngwako Pan Formation that is composed of continental clastic units (red beds); (iii) the D'Kar Formation, made up of shallow marine to lacustrine clastic sediments (sandstones and mudrocks) with minor oolitic limestone and rhythmites, and (iv) the Mamuno Formation that contains

Age	Group	Formation (thickness)	Lithology	Depositional setting
NEOPROTEROZOIC	GHANZI	MAMUNO (~ 1500 m)	Arkose, siltstone, shale, mudstone and limestone	Alluvial system changing to shallow marine and progradational shoreline
		D'KAR (~ 1500 m) Cu-Ag mineralization	Shale, mudstone, siltstone, arkose, marlstone and limestone	
		NGWAKO PAN (~ 2000 m)	Arkose, sub- arkose, siltstone and shale	
		KUKE (~ 500 m)	Sandstone and conglomerate	
MESOPROTEROZOIC	KGWEBE VOLCANICS (~ 2000 m)		Metarhyolite, tuff, metabasalt, agglomerate, minor intercalated metasedimentary rocks	Bi-modal volcanism

Fig. 2. Lithostratigraphic chart of the Kgwebe Volcanics and the Ghanzi Group units showing the stratigraphic position of the Cu-Ag mineralisation (modified from Modie, 1996).

predominantly sandstone (red beds) and mudstone interbedded with subordinate limestone horizons (Modie, 2000). The D'Kar Formation is the principal hosts of the Banana Zone Cu-Ag mineralisation, with however minor mineralisation occurrences in the Ngwako Pan Formation. During the Pan African Damaran Orogeny that spans from ~650 Ma (Schwartz et al., 1995) to 560–530 Ma (Porada and Berhorst, 2000), the rift basin underwent tectonic inversion and orogenic closure leading to the collision of the Congo and the Kalahari Cratons. This resulted in regional scale folding, faulting and lower greenschist facies metamorphism (Schwartz et al., 1995).

3. Sampling and analytical methods

Due to the extensive cover of Phanerozoic strata and scarcity of outcrops in the area, sampling was conducted on drill cores availed by Khoemacau Copper Mining. The samples were collected from five locations within the Banana Zone, including: (i) the Chalcocite Zone (CC), (ii) New Discovery (ND), (iii) North Limb (NL), (iv) North East Fold (NEF), and; (v) South Limb Definition (SLD) (Fig. 1b; Table 1).

Fifteen drill cores were investigated for their lithology and structural features. Twenty representative samples, including 10 samples from the Ngwako Pan sandstones and 10 samples from the D'Kar sandstones were selected for petrographic analyses. Standard petrographic polished thin sections were studied using both transmitted and reflected light microscopes at the Botswana University of Sciences and Technology (BIUST). Modal composition

data of the investigated samples are listed in Table 2.

Ten of the least weathered/altered samples of sandstones from Ngwako Pan and D'Kar Formations (five each) and five samples of mudrocks from D'Kar Formation (Table 3) were selected for geochemical investigations. The major element compositions were measured using X-ray fluorescence (XRF) on fusion beads at Council of Geoscience (COG), Pretoria, South Africa. The fusion beads were prepared by fusing 1 g of roasted sample and 10 g of flux consisting of 49.5% $\text{Li}_2\text{B}_4\text{O}_7$, 49.5% LiBO_2 and 0.50% LiI at 1150 °C. An in house amphibolite reference material (12/76) was used to determine the quality data and the analytical accuracy. The analytical precision was determined based on the duplicated results and was generally better than 5% for all elements.

Trace elements abundances were determined by inductively coupled plasma-mass spectrometry (ICP-MS) with a Perkin Elmer Elan 6000 at COG. For trace element analyses, 0.2 g of milled sample was digested in Savillex beakers with HCl , HNO_3 , HClO_4 and HF . The samples were dried and dissolved in HClO_4 . The drying process was repeated and then samples were re-dissolved in HNO_3 and H_2O_2 . For the REE analyses 0.5 g of sample was fused with 12:22 lithium tetraborate and lithium metaborate flux in Pt crucible and then dissolved in dilute HNO_3 . Standard samples GSD 09, GSD 10, GSD 11, GSR 1 and GSR 3 were used for the calibration of trace elements and REE. Accuracy of trace and REE element analyses was within 5%.

REE concentrations were normalised to Post Archean Average Australian Shale (PAAS, Taylor and McLennan, 1985) for the mudrocks and C1-chondrite (Taylor and McLennan, 1985) for the

Table 1
Ngwako Pan and D'Kar sample locations and descriptions.

Sample I.D.	Drill hole I.D.	Sample depth (m)	Subzone	Easting	Northing	Description	Formation
CS1	HA 526-D	147.23–147.42	North East Fold	689,712.90	7,676,881.00	Pink sandstone, medium grained and massive	Ngwako Pan
CS2	HA 526-D	165.00–165.30	North East Fold	689,712.90	7,676,881.00	Pink sandstone, coarse grained and laminated	
CS3	HA 520-D	172.90–173.00	Chalcocite	664,753.40	7,658,726.00	Pink sandstone, fine grained and massive	
CS4	HA 251-D	682.32–682.37	New Discovery	673,241.80	7,666,516.00	Grey-green sandstone and siltstone intercalations, medium grained and massive	
CS5	HA 17-D	256.73–256.80	North Limb	687,791.08	7,675,877.11	Pebbly pink sandstone, laminated	
CS6	HA 486-D	80.00–81.18	South Limb Definition	684,852.83	7,670,814.28	Pinkish grey massive sandstone, medium grained	
CS7	HA 248-D	435.12–435.18	New Discovery	672,844.50	7,666,031.00	Pink sandstone, medium grained and massive	
CS8	HA 500-D	303.15–303.26	North Limb	672,925.90	7,665,915.76	Pink sandstone, coarse grained and massive	
CS9	HA 38-D	165.73–171.00		683,385.32	7,672,800.89	Pinkish grey massive sandstone, fine to coarse grained	
CS10	HA 524-D	348.66–348.76		663,966.50	7,660,146.00	Pink sandstone, fine grained and massive	
MS1	HA 515-D	68.29–68.37	Chalcocite	666,372.60	7,658,880.00	Grey sandstone with marl intercalations, medium grained and massive	D'Kar
MS2	HA 526-D	108.20–108.37	North East Fold	689,712.90	7,676,881.00	Red sandstone, fine to medium grained and massive; red coating on the grains margin	
MS3	HA 520-D	108.17–108.31	Chalcocite	664,753.40	7,658,726.00	Pink sandstone, fine grained and massive	
MS4	HA 251-D	161.85–162.00	New Discovery	673,241.80	7,666,516.00	Grey sandstone, coarse grained and massive	
MS5	HA 17-D	206.45–206.64	North Limb	687,791.08	7,675,877.11	Grey-green sandstone and siltstone intercalations	
MS6	HA 10-D	138.37–138.58	North East Fold	689,030.01	7,676,533.74	Grey-green sandstone, fine grained and massive	
MS7	HA 10-D	169.27–169.40	North East Fold	689,030.01	7,676,533.74	Grey sandstone with marl intercalations, coarse grained and massive	
MS8	HA 430-D	127.70–127.92	South Limb Definition	686,152.54	7,672,042.48	Grey sandstone interbedded with siltstone, medium grained and laminated	
MS9	HA 430-D	164.30–164.50	North Limb	686,152.54	7,672,042.48	Grey sandstone with siltstone and limestone intercalations, medium grained and laminated	
MS10	HA 38-D	144.30–144.40		683,385.32	7,672,800.89	Grey sandstone, fine grained, laminated	

Table 2

Detrital modal compositions of selected sandstone samples from the Ghanzi Group.

Sample No.	Qt	F	L	Qm	F	Lt	Qp	Lm	Lv	Ls	Qm	P	K
CS1	68.9	22.7	8.4	68.9	22.7	8.4	20.0	0.0	0.0	80.0	73.8	9.2	17.0
CS2	57.9	36.8	5.3	55.4	36.8	7.8	66.7	0.0	0.0	33.3	58.7	16.3	25.0
CS3	58.6	40.4	2.0	52.5	40.4	7.1	77.8	0.0	0.0	22.2	55.6	16.7	27.7
CS4	65.7	34.3	1.0	64.6	34.2	2.0	50.0	0.0	0.0	50.0	65.9	10.1	24.0
CS5	76.0	19.0	5.0	73.0	19.0	8.0	78.8	9.1	6.1	6.1	75.0	11.9	13.1
CS6	65.0	35.0	0.0	60.0	35.0	5.0	100.0	0.0	0.0	0.0	63.2	15.7	21.1
CS7	66.3	31.6	2.1	63.2	31.6	5.2	60.0	20.0	0.0	20.0	66.7	14.4	18.9
CS8	58.6	34.3	7.1	56.5	34.3	9.2	45.0	10.0	10.0	10.0	60.5	14.2	25.3
CS9	64.0	28.0	8.0	64.0	28.0	8.0	52.4	9.5	0.0	38.1	67.8	11.4	20.8
CS10	56.8	42.1	1.1	54.7	42.1	3.2	66.7	0.0	0.0	33.3	56.5	18.5	25.0
Average	63.8	32.4	4.0	61.3	32.4	6.4	61.7	4.9	1.6	29.3	64.4	13.8	21.8
MS1	79.6	20.4	0.0	77.6	20.4	2.0	100.0	0.0	0.0	0.0	79.2	8.4	12.4
MS2	58.6	39.4	2.0	55.6	39.4	5.0	60.0	0.0	0.0	40.0	58.5	26.6	14.9
MS3	50.0	43.9	6.1	48.9	43.9	7.2	30.0	20.0	0.0	50.0	51.1	17.0	31.9
MS4	54.0	38.8	7.2	51.9	38.8	9.3	40.0	60.0	0.0	0.0	55.8	15.7	28.5
MS5	75.0	22.0	3.0	70.0	22.0	8.0	62.5	12.5	0.0	25.0	76.1	6.5	17.4
MS6	65.7	30.3	4.0	63.6	30.3	6.1	33.3	16.7	0.0	50.0	67.7	14.0	18.3
MS7	64.3	30.6	5.1	59.2	30.6	10.2	50.0	20.0	0.0	30.0	65.9	13.6	20.5
MS8	75.0	25.0	0.0	75.0	25.0	0.0	0.0	0.0	0.0	0.0	75.0	15.0	10.0
MS9	70.0	30.0	0.0	67.0	30.0	3.0	100.0	0.0	0.0	0.0	69.1	12.3	18.6
MS10	75.0	22.5	2.5	75.0	22.5	2.5	0.0	0.0	0.0	100.0	76.9	7.7	15.4
Average	66.7	30.3	3.0	64.4	30.3	5.3	47.6	12.9	0.0	29.5	67.5	13.7	18.8

CS = Ngwako Pan sandstones; MS = D'Kar sandstones; Qt = Total quartz (Qm + Qp); Qm = Monocrystalline quartz; Qp = Polycrystalline quartz; F = Total feldspar grains (P + K); P = Plagioclase grains; K = K-feldspar grains; L = Unstable lithic fragments (Lm + Lv + Ls); Lm = Metamorphic lithic fragments; Lv = Volcanic fragments; Ls = Sedimentary fragments; Lt = Total lithic fragments (L + Qp). After Dickinson (1985).

sandstones. For PAAS-normalised samples, REE fractionation was calculated as Pr_{SN}/Yb_{SN} (SN, shale normalised) avoiding anomalous La and Ce concentrations (Franchi et al., 2016). In order to avoid anomalous behaviour of La, Ce, Eu and giving the quasi-linear behaviour of the near neighbour REE in the logarithmic plots, REE anomalies were calculated using the geometric equation of Lawrence et al. (2006). In chondrite normalisation, Eu anomaly was calculated as $Eu/Eu^* = Eu_{CN}/((Sm_{CN} \cdot Gd_{CN})^{1/2})$ (Taylor and McLennan, 1985). The subscript "CN" indicates Chondrite-normalised values. The Ce anomaly was defined as $Ce/Ce^* = Ce_{CN}/((La_{CN} \cdot Pr_{CN})^{1/2})$ (Kato and Nakamura, 2003).

4. Results

4.1. Petrography

4.1.1. Ngwako Pan Formation

Sandstones of the Ngwako Pan Formation vary from poorly- to moderately sorted, consisting of fine sand size clasts to very coarse sandstone and micro-rudite (0.125–2.30 mm in size). These sandstones are texturally immature to submature. Finer grains are generally angular to sub angular, whereas the medium to coarse-grained clasts are subrounded to well-rounded. The studied samples are mainly composed of quartz (Qt), feldspar (F) and lithic fragments (L). The average compositional ratio is $Qt_{64}: F_{32}: L_4$, and falls within the arkose field (Fig. 3).

Quartz is the predominant framework grain and occurs as both mono- and polycrystalline quartz. Monocrystalline quartz is the most common type ranging from 52.5 to 73.0%, whereas polycrystalline quartz is present in a relatively small amount ranging from 0.0 to 6.1% (average 2.5%) (Table 2). Most of the monocrystalline quartz grains are inclusions-free and characterised by non-undulose extinction (Fig. 4a). Locally quartz grains in the medium to coarse sand size range exhibit undulose extinction. Four types of polycrystalline quartz grains were observed: (i) grains composed of few equant crystals with no or little intercrystalline suturing (Fig. 4a); (ii) grains composed of crystals with both sharp and sutured boundaries (Fig. 4b); (iii) grains composed of polygonal

crystals intersecting at an angle of 120° (Fig. 4c) and (iv) recrystallised metamorphic quartz grains (Fig. 4d).

The studied sandstones are rich in feldspar (19.0–42.1%; average 32.4%, Table 2), which is either unaltered or heavily weathered and altered. Different stages of argillisation and carbonatisation were observed. K-feldspar (average 19.2%) is the most abundant feldspar and it includes microcline (Fig. 4c), orthoclase and microperthite (Fig. 4e and f). Muscovite, heavy minerals (magnetite, zircon and rutile) and myrmekite occur in minor quantities, but are invariably conspicuous and important framework components.

Lithoclasts are dominated by mudstone fragments. Other lithoclasts found include recrystallised chert, quartzite, schist, and volcanic lithics (Fig. 5).

The amount of matrix is highly variable and is always below 15% modal. The matrix is composed of phyllosilicates (chlorite, sericite and muscovite) and silt size clasts of quartz and feldspar. In addition to the matrix, the grains are cemented by mostly sparry calcite and locally by quartz that grows in crystallographic and optical continuity with the quartz clasts. Late iron oxide grain-coating cement is also present.

4.1.2. D'Kar Formation

Sandstones of the D'Kar Formation are poorly- to moderately sorted (grains size ranging between 0.08 and 0.75 mm) and texturally immature to submature. These sandstones are characterised by grains of variable roundness similar to the Ngwako Pan sandstones. The sandstones from the D'Kar Formation have an average compositional ratio of $Qt_{67}: F_{30}: L_3$ (Table 2) and are classified as subarkoses, and arkoses (Fig. 3).

Monocrystalline inclusion-free quartz with non-undulose extinction (Fig. 4a) is the most common quartz type ranging from 48.9 to 77.6% (average 64.4%, Table 2). Few grains with undulose extinction are also observed. Polycrystalline quartz occurs in a relatively small amount (average 2.3%, Table 2). K-feldspar clasts (average 17.4%) including microcline, orthoclase and microperthite (Fig. 4c, e, f), dominate over plagioclase (average 12.9%) (Fig. 4a). Feldspar has undergone variable stages of alteration, including, argillisation, sericitisation and carbonatisation. Lithic fragments

Table 3

Major, minor, and trace element concentrations, major oxide ratios and Chemical Index of Weathering (CIW') of the Ngwako Pan sandstones, D'Kar sandstones and D'Kar mudrocks.

	Ngwako Pan Sandstones						D'Kar Sandstones						D'Kar mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg	MS1	MS2	MS3	MS4	MS5	Avg	MD1	MD2	MD3	MD4	MD5	Avg
Major and minor elements (wt. %)																		
SiO ₂	73.32	75.32	77.18	75.99	76.58	75.68	64.00	75.45	74.73	75.91	75.68	73.15	63.89	64.15	61.71	61.85	60.48	62.42
TiO ₂	0.20	0.39	0.29	0.30	0.29	0.29	0.24	0.46	0.35	0.42	0.45	0.38	0.79	0.75	0.65	0.73	0.86	0.76
Al ₂ O ₃	7.41	9.54	8.43	8.55	8.49	8.48	7.03	9.93	9.48	9.61	9.77	9.17	15.88	14.48	12.06	15.56	17.67	15.13
Fe ₂ O ₃ (t)	0.91	1.92	1.53	1.47	1.50	1.46	0.92	2.38	1.65	2.16	2.27	1.88	5.10	5.55	5.47	6.63	5.34	5.62
MnO	0.16	0.09	0.09	0.11	0.10	0.11	0.10	0.08	0.09	0.10	0.09	0.09	0.11	0.14	0.24	0.12	0.11	0.14
MgO	0.35	0.72	0.41	0.50	0.45	0.49	0.48	0.69	0.59	0.52	0.60	0.57	2.82	2.64	3.86	2.96	3.33	3.12
CaO	6.96	2.84	3.66	4.55	4.10	4.42	5.07	2.51	3.79	3.64	3.08	3.62	1.54	2.46	5.37	2.48	1.46	2.66
Na ₂ O	1.76	2.57	2.29	2.23	2.26	2.22	1.76	2.97	2.37	2.26	2.62	2.39	1.96	2.8	2.84	1.9	2.84	2.47
K ₂ O	3.42	3.35	2.98	3.29	3.13	3.23	2.82	2.92	3.50	2.85	2.88	2.99	4.45	3.02	1.63	3.71	4.35	3.43
P ₂ O ₅	0.06	0.09	0.07	0.07	0.07	0.07	0.06	0.11	0.09	0.07	0.09	0.08	0.19	0.18	0.17	0.18	0.21	0.19
Cr ₂ O ₃	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01
LOI	5.03	2.21	2.76	2.38	2.27	2.93	16.74	2.16	2.82	2.31	2.24	5.25	2.8	3.35	5.65	3.42	2.93	3.63
Total	99.59	99.05	99.7	99.44	99.27	99.41	99.23	99.67	99.45	99.86	99.77	99.60	99.54	99.53	99.66	99.56	99.59	99.58
Na ₂ O/K ₂ O	0.51	0.77	0.77	0.68	0.72	0.69	0.62	1.02	0.68	0.79	0.91	0.80	0.44	0.93	1.74	0.51	0.65	0.85
SiO ₂ /Al ₂ O ₃	9.89	7.90	9.16	8.89	9.02	8.97	9.10	7.60	7.88	7.90	7.74	8.05	4.02	4.43	5.12	3.97	3.42	4.19
Al ₂ O ₃ /TiO ₂	37.05	24.46	29.07	28.84	28.95	29.67	29.29	21.59	27.09	21.36	23.27	24.52	20.10	19.31	18.55	21.32	20.55	19.96
Fe ₂ O ₃ +MgO	1.26	2.64	1.94	1.97	1.95	1.95	1.40	3.07	2.24	2.68	2.87	2.45	7.92	8.19	9.33	9.59	8.67	8.74
CIW'	81.00	79.00	79.00	79.00	79.00	79.00	80.00	77.00	80.00	81.00	79.00	79.00	89.00	84.00	81.00	89.00	86.00	86.00
Li	6.04	19.78	14.50	13.55	14.03	13.58	6.22	12.75	9.49	10.31	11.53	10.06	37.35	36.27	28.32	50.14	46.71	39.76
Be	0.79	1.30	1.56	1.22	1.39	1.25	1.16	0.66	0.91	1.08	0.87	0.94	4.10	2.83	1.58	3.70	4.49	3.34
V	29.25	53.95	39.72	41.41	40.56	40.98	33.75	54.97	44.36	47.29	51.13	46.30	115.6	106.17	98.11	119.28	130.00	113.83
Cr	22.49	37.66	22.3	27.86	25.08	27.08	15.47	27.59	21.53	22.49	25.04	22.42	71.34	64.21	58.42	63.96	72.13	66.01
Co	10.05	8.78	6.77	8.65	7.71	8.39	3.45	7.72	5.59	6.16	6.94	5.97	24.00	15.90	37.64	16.41	36.50	26.09
Ni	11.85	19.83	22.21	18.06	20.13	18.42	9.00	16.53	12.77	13.26	14.89	13.29	27.61	29.36	32.52	23.56	24.15	27.44
Cu	4108.20	1871.60	2499.60	2861.29	2680.46	2804.24	1004.6	37.55	521.07	843.42	440.48	569.42	5705.40	4205.40	2274.20	648.26	4914.60	3549.57
Zn	22.89	82.86	64.75	57.24	61.00	57.75	22.11	776.09	399.10	147.77	461.93	361.40	121.33	162.02	237.58	96.89	174.77	158.52
Ga	7.44	12.19	10.60	10.17	10.38	10.16	8.13	11.83	9.98	8.75	10.29	9.80	27.19	23.36	18.79	25.39	29.01	24.75
Rb	84.67	107.68	97.76	97.65	97.70	97.09	75.35	66.19	70.77	73.82	70.01	71.23	263.95	120.26	51.89	177.27	205.56	163.79
Sr	60.91	41.73	54.66	52.91	53.79	52.80	58.50	48.55	53.53	56.84	52.70	54.02	35.68	58.37	63.77	39.61	38.68	47.22
Mo	1.15	51.90	2.73	19.10	10.91	17.16	12.62	0.49	6.56	10.60	5.54	7.16	8.12	2.37	29.05	5.77	3.88	9.84
Ag	<0.50	<0.50	<0.50	<0.50	<0.50	-	<0.50	<0.50	<0.50	<0.50	<0.50	-	<0.50	<0.50	<0.50	<0.50	<0.50	-
Cd	0.18	0.24	0.14	0.19	0.16	0.18	0.16	8.01	4.09	1.47	4.74	3.69	0.08	0.10	0.26	0.14	0.21	0.16
Te	<0.01	<0.01	<0.01	<0.01	<0.01	-	<0.01	<0.01	<0.01	<0.01	<0.01	-	<0.01	0.02	0.02	0.01	<0.01	0.02
Ba	787.63	749.79	731.07	764.23	747.65	756.07	733.47	822.45	777.96	748.30	785.38	773.51	822.27	609.91	323.44	614.25	944.82	662.94
Tl	0.46	0.54	0.56	0.52	0.54	0.53	0.41	0.39	0.40	0.41	0.40	0.40	1.05	0.54	0.27	0.83	1.21	0.78
Pb	43.62	9.50	14.75	23.01	18.88	21.95	5.56	513.06	259.31	90.14	301.60	233.94	4.16	6.90	7.70	72.62	4.39	19.15
Bi	67.59	2.95	3.77	25.44	14.60	22.87	1.42	0.16	0.79	1.21	0.69	0.85	4.14	0.41	2.85	0.54	1.31	1.85
U	2.50	2.03	1.35	1.99	1.67	1.91	1.77	1.34	1.56	1.70	1.52	1.58	3.78	3.98	6.75	4.07	4.84	4.68
Th	4.40	6.85	11.64	7.63	9.63	8.03	4.63	4.70	4.67	4.84	4.65	4.70	6.18	9.70	9.32	11.37	11.07	9.53
Zr	86.19	168.26	183.41	146.66	165.04	149.91	117.22	180.70	149.96	147.80	144.25	147.99	153.84	217.76	236.46	186.16	189.98	196.84

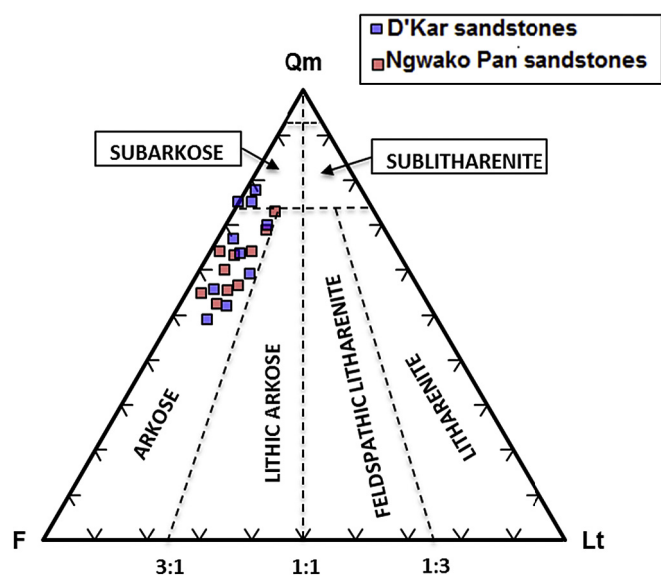


Fig. 3. Ternary plot for classification of sandstones (after Folk, 1980). Qm = Monocrystalline quartz, F = Total feldspar grains (Plagioclase + K-feldspar), Lt = Total siliciclastic lithic fragments.

range from 0.0 to 10.2% and are largely composed of mudstone (Table 2). Metamorphic rock fragments and granitic rock fragments are subordinate (Table 2, Fig. 5).

The matrix proportion of D'Kar sandstone is highly variable (1.0–12.0%). The matrix is composed of argillaceous materials (detrital clay, sericite and chlorite). The grains are cemented by sparry calcite and quartz overgrowths in optical continuity with quartz grains.

Mudrocks of the D'Kar Formation are composed of green to grey siltstones and mudstones. The siltstones are very fine grained and finely laminated. They consist essentially of silty quartz and subordinate feldspar, aligned muscovite, sericite and chlorite flakes in a calcareous matrix (Fig. 6a). The mudstones are finely laminated and consist of elliptical quartz in a muscovite and chlorite-rich matrix (Fig. 6b). The quartz and chlorite grains have their long axes aligned parallel to the lamination. Locally, the mudstones contain high amounts of organic matter, giving the rock its black colour.

5. Geochemical composition of sandstones and mudrocks

5.1. Major elements

Major element oxides and trace element concentrations are given in Table 3 along with major oxide ratios.

5.1.1. Ngwako Pan Formation

Sandstones of the Ngwako Pan Formation have a wide range in SiO_2 contents varying from 64.00 to 75.90 wt% and a restricted range in Al_2O_3 contents varying from 7.03 to 9.93 wt %. These sandstones are characterised by TiO_2 contents of 0.20–0.39 wt%, K_2O contents of 2.98–3.42 wt% and Na_2O contents of 1.76–2.57 wt %. The Fe_2O_3 , MgO and CaO contents vary from 0.91 to 1.92 wt %, from 0.35 to 0.72 wt % and from 2.84 to 6.96 wt %, respectively. The $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios for these sandstones range from 7.90 to 9.89 and from 24.46 to 37.05, respectively. The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ values are between 0.51 and 0.77.

5.1.2. D'Kar Formation

Sandstones of the D'Kar Formation have a restricted range of

SiO_2 and Al_2O_3 contents varying from 73.32 to 77.18 wt % and from 7.41 to 9.54 wt %, respectively. These sandstones show slightly higher contents of TiO_2 , K_2O and Na_2O (0.24–0.46 wt %, 2.82–3.50 wt % and 1.76–2.97 wt %, respectively) with respect to sandstones of the Ngwako Pan Formation. The Fe_2O_3 and MgO contents vary from 0.92 to 2.38 wt % and from 0.48 to 0.69 wt %, respectively. The CaO contents in these samples vary from 2.51 to 5.07 wt %. The $\text{Al}_2\text{O}_3/\text{TiO}_2$ and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios in these samples range from 21.36 to 29.29 and from 0.62 to 1.02, respectively. The $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios range from 7.60 to 9.10.

Mudrocks of the D'Kar Formation are characterised by lower SiO_2 (60.48–64.15 wt %), but relatively higher Al_2O_3 (12.6–17.67 wt %) and TiO_2 (0.65–0.86 wt %) contents with respect to the D'Kar sandstones. The K_2O and Na_2O contents vary from 1.64 to 4.45 wt % and from 1.90 to 2.84 wt %, respectively. Mudrocks are richer in Fe_2O_3 (5.1–6.63 wt %) and MgO (2.64–3.86 wt %) in comparison to the D'Kar sandstones. The CaO contents range from 1.46 to 5.37 wt %. The $\text{SiO}_2/\text{Al}_2\text{O}_3$, $\text{Al}_2\text{O}_3/\text{TiO}_2$ and $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios range from 3.42 to 5.19, from 18.55 to 21.32 and from 0.44 to 1.74, respectively.

5.2. Trace elements

Trace element contents of the investigated samples are shown in Table 3. The next paragraphs present the distribution of selected trace elements in the Ngwako and D'Kar Formations.

5.2.1. Ngwako Pan Formation

The Cr, Co and Ni contents in the sandstones of the Ngwako Pan Formation range from 22.30 to 37.66 ppm, 6.77–10.05 ppm and 11.85–22.21 ppm, respectively. The Th and Zr contents range from 4.40 to 11.64 ppm and from 86.19 ppm to 183.41 ppm, respectively. The concentration of V in these sandstones is between 29.25 and 53.95 ppm. The La/Co and La/Cr values range from 1.02 to 5.84 and 0.46 to 1.77, respectively. The Th/Cr and Th/Co ratios in these sandstones range from 0.18 to 0.52, respectively (Table 5). Furthermore, these sandstones have an average Cr/Th ratio of 3.76 (Table 5).

5.2.2. D'Kar Formation

The Cr, Co and Ni contents in the sandstones of the D'Kar Formation range from 15.47 to 27.59 ppm, from 3.45 to 7.72 ppm and from 9.00 to 16.53 ppm, respectively. The Th contents ranging from 4.63 to 4.83 ppm and Zr contents ranging from 117.22 to 180.70 ppm. The V content varies from 33.75 to 54.97 ppm. The La/Co and La/Cr ratios range from 2.82 to 5.24 and 0.79 to 1.17, respectively. The Th/Cr and Th/Co values in these sandstones vary from 0.19 to 0.30 and from 0.61 to 1.34, respectively (Table 5). The Cr/Th ratio has an average value of 4.77 (Table 5).

In comparison with the sandstone samples, the mudrocks of the D'Kar Formation show higher content of Cr, Co and Ni (63.96–72.13 ppm, 15.90–37.64 ppm and 24.15–32.52 ppm, respectively). Thorium contents in these mudrocks vary from 6.18 to 11.37 ppm. They are characterised by relatively high Zr and V contents (153.84 ppm–236.46 ppm and 98.11–130.00 ppm, respectively). The La/Co and La/Cr values range from 0.80 to 2.62 and 0.27 to 0.70, respectively, while the Th/Cr and Th/Co ratios vary from 0.09 to 0.18 and 0.25 to 0.69, respectively (Table 5). These samples are characterised by high Cr/Th ratio with an average value of 7.31 (Table 5).

5.3. Rare earth elements

The REE data are presented in Table 4 and are shown as chondrite-normalised patterns in Fig. 7a and b for sandstones and as PAAS-normalised patterns in Fig. 7c for mudrocks. The total REE

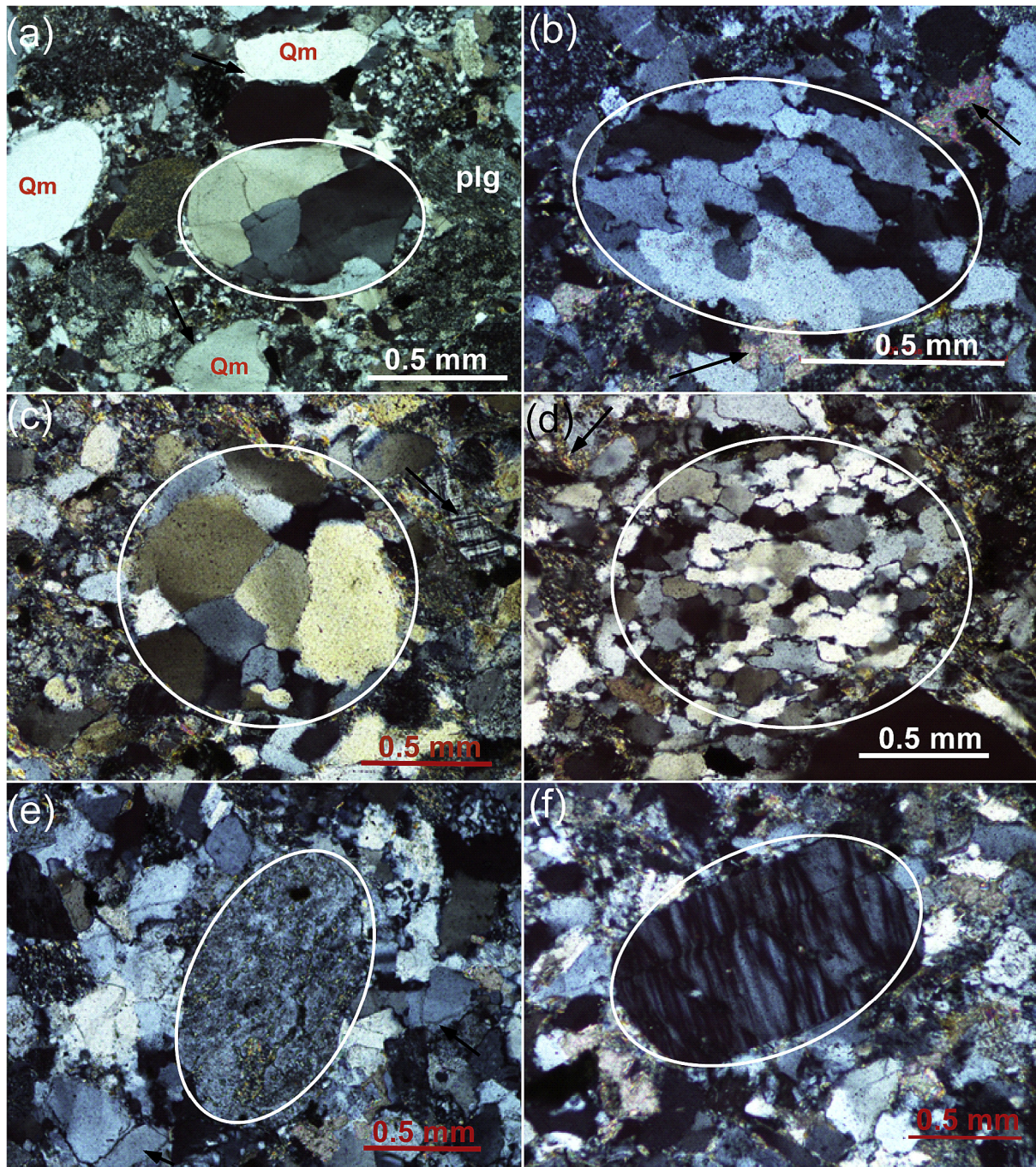


Fig. 4. Photomicrographs (transmitted light) of different types of polycrystalline quartz grains (white circles), monocrystalline quartz grains, feldspar grains and different cements. All photomicrographs are in cross polarized light. (a) Polycrystalline quartz of probable igneous origin; the grain is composed of few equant crystals with little or no intercrystalline suturing; quartz cement (arrow); monocrystalline quartz (Qm) grains with non-undulose extinction and partially altered plagioclase (plg) grain. (b) Polycrystalline quartz of probable metamorphic origin; the grain consists of crystals with both sharp and sutured boundaries; calcite cement (arrows). (c) Polycrystalline quartz grains of probable metamorphic origin with polygonal crystal boundaries at 120°; Microcline (arrow). (d) Polycrystalline quartz grain of probable metamorphic origin with clear evidences of recrystallisation; Clay minerals (arrow). (e) Partially altered orthoclase grain; quartz cement (arrow). (f) Micropertite grain.

contents (ΣREE) of the investigated samples range from 56.1 to 211.2 ppm (average 137.6 ppm) in the sandstones of the Ngwako Pan Formation, from 91.2 to 123.0 ppm (average 111.3 ppm) in the sandstones of the D'Kar Formation and from 105.3 to 210.9 ppm (average 187.6 ppm) in the mudrocks of the D'Kar Formation.

5.3.1. Ngwako Pan Formation

When normalised to chondrite, the sandstones of the Ngwako

Pan Formation (Fig. 7a) reveal an enriched LREE ($\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}} = 3.27\text{--}2.68$; average 2.95 ± 0.21) and a flat HREE ($\text{Gd}_{\text{CN}}/\text{Yb}_{\text{CN}} = 1.41\text{--}1.09$; average 1.30 ± 0.13) pattern that is very similar to PAAS except for low $\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}}$ ratios. These samples show a variable Eu negative anomaly ($\text{Eu}/\text{Eu}^* = 0.61\text{--}0.83$; average 0.70 ± 0.09) and a slightly negative Ce anomaly ($\text{Ce}/\text{Ce}^* = 0.91\text{--}0.93$; average 0.92 ± 0.01).

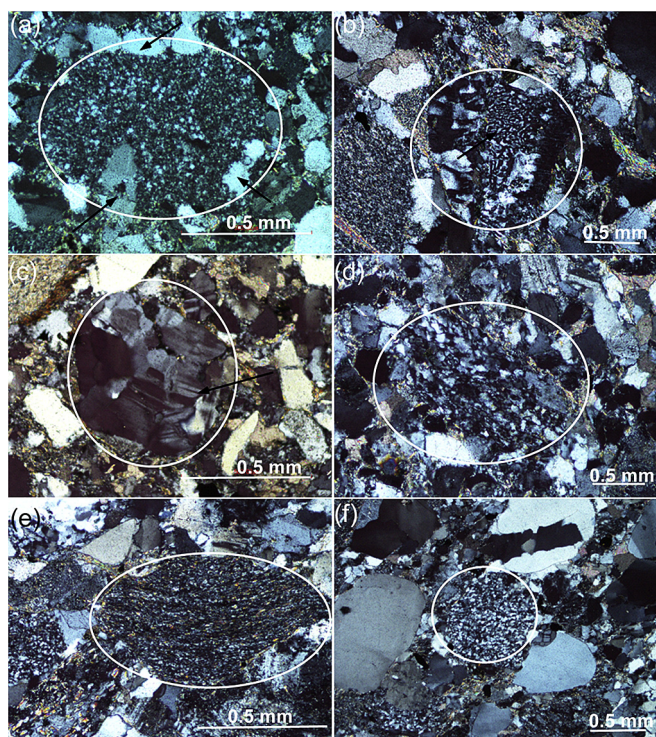


Fig. 5. Photomicrographs of lithic grains (white circles). All photomicrographs are in cross polarized light. (a) Volcanic rock fragments containing K-feldspar grains (arrows). (b, c) Fragments of a granitoid rock; myrmekitic intergrowth (arrow) can be recognized in (b) and twin lamellae (arrow) in (c). (d) Fragment of quartzite. (e) Mudrock fragment (f) Recrystallised chert fragment.

5.3.2. D'Kar Formation

The chondrite normalised REE pattern for the sandstones of the D'Kar Formation is very similar to PAAS except for low $\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}}$ ratios (Fig. 7b). The samples show LREE enrichment ($\text{La}_{\text{CN}}/\text{Sm}_{\text{CN}} = 3.61\text{--}2.68$; average 3.13 ± 0.42) coupled with flat HREE profile ($\text{Gd}_{\text{CN}}/\text{Yb}_{\text{CN}} = 1.30\text{--}1.11$; average 1.19 ± 0.07) and a variable negative Eu anomaly ($\text{Eu}/\text{Eu}^* = 0.69\text{--}0.81$; average 0.75 ± 0.05). The Ce anomaly is slightly negative ($\text{Ce}/\text{Ce}^* = 0.83\text{--}0.91$; average 0.90 ± 0.04).

All but one sample (MD1) of the mudrocks of the D'Kar Formation show an overall flat PAAS-normalised REE pattern (average $\text{Pr}_{\text{SN}}/\text{Yb}_{\text{SN}} = 0.84 \pm 0.11$ without MD1) (Fig. 7c). The mudrocks are characterised by variable Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.81\text{--}1.21$; average 0.93 ± 0.16), a slightly negative Ce anomaly ($\text{Ce}/\text{Ce}^* = 0.86\text{--}0.90$; average 0.87 ± 0.05), and Y/Ho ratios between 26.27 and 28.80 (average 27.85).

6. Discussion

6.1. Sandstone classification

Sandstone classification followed in this study is based on the matrix content and mineralogical composition (Dott, 1964; Pettijohn et al., 1973, 1987; Folk, 1974, 1980). Given that the matrix content of the investigated sandstones is $\leq 15\%$ modal and their Al_2O_3 contents is relatively low (average 8.48 and 9.17 wt% in the Ngwako Pan and D'Kar Formations, respectively), the sandstones of both the Ngwako Pan and D'Kar Formations are classified as arenites. The relative proportions of the three major framework grains, namely monocrystalline quartz (Qm), total feldspar (F) and total lithic fragments (Lt), suggests that the sandstones of the Ngwako

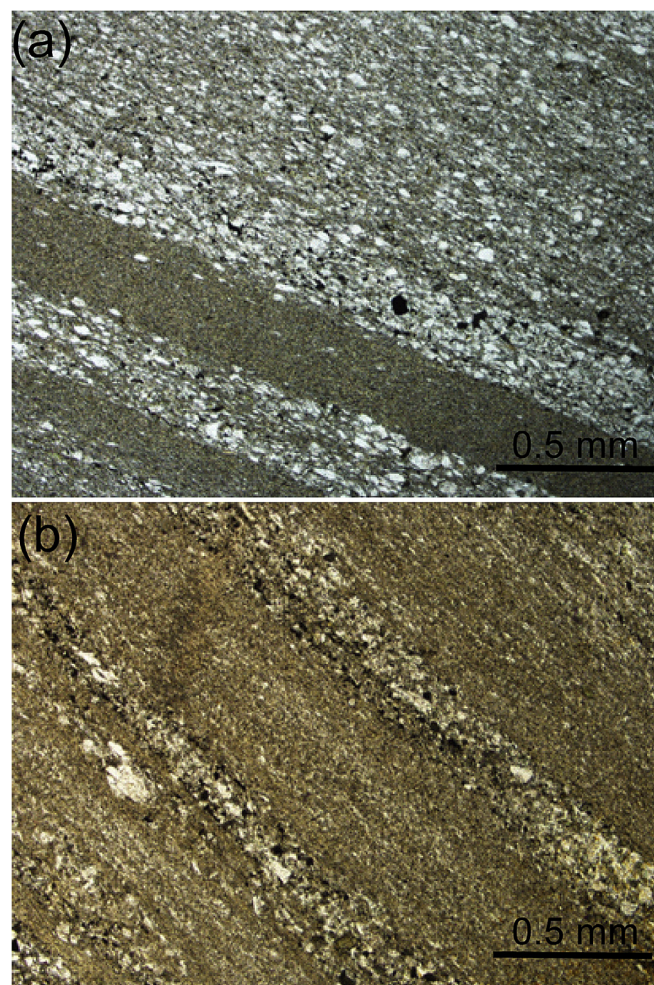


Fig. 6. Photomicrographs (transmitted light) of mudrocks of D'Kar Formation. (a) Siltstone showing fine laminations. (b) Mudstone showing fine lamination.

Pan Formation can be further classified as arkose, whereas the sandstones of the D'Kar Formation are classified as subarkose, and arkose (Fig. 3). Based on the geochemical classifications proposed by Herron (1988) sandstones from both D'kar and Ngwako Pan Formations are classified as arkose (Fig. 8a), (see also Lindsey, 1999). These sandstones can be classified as potassic following the classification of Blatt et al. (1972) (Fig. 8b). Sandstones of the Ngwako Pan and D'kar Formations are poorly- to moderately-sorted and the roundness ratio between fine- and coarse-grained is correlated with the size of the clasts pointing towards texturally immature to submature sediments (Folk, 1951). Mineralogical maturity can be deduced from $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios that reflect the abundance of quartz, feldspar and clay contents within the sediments (Potter, 1978). Low average values of $\text{SiO}_2/\text{Al}_2\text{O}_3$ found in the studied samples (8.97 for the sandstones of the Ngwako Pan Formation and 8.05 for the sandstones of the D'Kar Formation) are indicative of low maturity (Table 3). The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios can also be used in determining chemical maturity as the $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio reflects the liable nature of plagioclase relative to K-feldspar (Pettijohn et al., 1987). The $\text{Na}_2\text{O}/\text{K}_2\text{O}$ average values of 0.69 for the sandstones of the Ngwako Pan Formation and 0.80 for the sandstones of the D'Kar Formation indicate that the samples are chemically immature (Table 3).

Table 4Rare earth elements and Yttrium (REE + Y) abundance (ppm), Σ REE, Y/Ho, REE ratios and Eu and Ce anomalies of the Ngwako Pan sandstones, D'Kar sandstones and D'Kar mudrocks.

	Ngwako Pan Sandstones						D'Kar Sandstones						D'Kar mudrocks					
	CS1	CS2	CS3	CS4	CS5	Avg.	MS1	MS2	MS3	MS4	MS5	Avg.	MD1	MD2	MD3	MD4	MD5	Avg.
Y	13.57	21.49	40.90	25.26	33.08	26.86	18.38	25.94	23.16	26.64	25.79	23.98	25.06	36.78	31.85	38.17	40.41	34.45
La	10.26	23.51	39.52	24.37	31.95	25.92	18.08	21.77	20.93	23.70	24.83	21.86	19.15	41.67	33.50	35.55	50.56	36.09
Ce	20.22	47.58	80.54	49.32	64.93	52.52	35.10	42.21	39.66	44.29	45.75	41.40	38.51	81.27	67.39	72.29	101.78	72.25
Pr	2.66	6.12	10.57	6.43	8.50	6.86	4.48	5.60	6.04	6.17	5.73	5.60	5.01	10.73	8.95	9.42	13.24	9.47
Nd	10.72	23.42	41.74	25.22	33.48	26.92	17.53	21.92	20.73	24.26	24.59	21.81	20.49	41.59	34.79	37.11	50.98	36.99
Sm	2.41	4.52	8.56	5.15	6.85	5.50	3.22	4.60	4.91	5.25	4.33	4.46	4.55	8.09	7.09	7.78	9.68	7.44
Eu	0.62	1.05	1.62	1.10	1.36	1.15	0.78	1.00	1.03	1.08	1.11	1.00	1.14	1.44	1.15	1.42	1.63	1.36
Gd	2.15	4.03	7.74	4.62	6.18	4.95	2.97	4.3	3.64	3.99	4.05	3.79	4.18	7.21	6.16	6.81	8.47	6.57
Tb	0.40	0.71	1.39	0.83	1.11	0.89	0.54	0.79	0.67	0.68	0.74	0.68	0.77	1.24	1.07	1.21	1.45	1.15
Dy	2.43	4.07	7.83	4.76	6.30	5.08	3.13	4.63	3.88	4.38	4.58	4.12	4.52	6.96	5.98	6.84	7.86	6.43
Ho	0.50	0.78	1.52	0.93	1.23	0.99	0.64	0.90	0.77	0.85	0.91	0.81	0.87	1.35	1.14	1.34	1.51	1.24
Er	1.58	2.43	4.37	2.79	3.58	2.95	1.98	2.73	2.36	2.42	2.77	2.45	2.64	3.99	3.51	4.04	4.40	3.72
Tm	0.25	0.38	0.68	0.44	0.56	0.46	0.32	0.41	0.38	0.42	0.41	0.39	0.42	0.63	0.52	0.62	0.69	0.58
Yb	1.60	2.47	4.45	2.84	3.64	3.00	2.06	2.68	2.66	2.71	2.77	2.58	2.65	4.11	3.46	3.97	4.46	3.73
Lu	0.27	0.37	0.68	0.44	0.56	0.46	0.33	0.39	0.38	0.43	0.42	0.39	0.42	0.65	0.54	0.61	0.65	0.57
Σ REE	56.07	121.44	211.21	129.24	170.22	137.64	91.16	113.93	108.01	120.62	122.99	111.34	105.30	210.90	175.30	189.00	257.40	187.57
(Σ REE)	46.27	105.15	180.93	110.49	145.71	117.71	78.41	96.10	92.27	103.67	105.23	95.14	87.71	183.35	151.72	162.15	226.24	162.23
(Σ HREE)	9.18	15.24	28.66	17.65	23.16	18.78	11.97	16.83	14.74	15.88	16.65	15.21	16.47	26.14	22.38	25.44	29.49	23.98
LREE/HREE	5.04	6.90	6.31	6.26	6.29	6.16	6.55	5.71	6.26	6.53	6.32	6.27	5.33	7.01	6.78	6.37	7.67	6.63
(Eu/Eu*) _{CN}	0.83	0.75	0.61	0.69	0.64	0.70	0.77	0.69	0.74	0.72	0.81	0.75	-	-	-	-	-	-
(Eu/Eu*) _{SN}	-	-	-	-	-	-	-	-	-	-	-	-	1.21	0.88	0.81	0.90	0.84	0.93
(Ce/Ce*) _{CN}	0.91	0.93	0.92	0.92	0.92	0.92	0.91	0.90	0.83	0.86	0.90	0.90	-	-	-	-	-	-
(Ce/Ce*) _{SN}	-	-	-	-	-	-	-	-	-	-	-	-	0.97	0.86	0.86	0.90	0.86	0.89
(La/La*) _{SN}	-	-	-	-	-	-	-	-	-	-	-	-	1.00	0.92	0.89	0.92	0.89	0.92
La _{CN} /Sm _{CN}	2.68	3.27	2.91	2.98	2.94	2.95	3.53	2.98	2.68	2.84	3.61	3.13	-	-	-	-	-	-
Gd _{CN} /Yb _{CN}	1.09	1.32	1.41	1.32	1.38	1.30	1.17	1.30	1.11	1.19	1.18	1.19	-	-	-	-	-	-
(Pr/Yb) _{SN}	-	-	-	-	-	-	-	-	-	-	-	-	0.60	0.83	0.83	0.76	0.95	0.79
Y/Ho	27.14	27.55	26.91	27.16	26.89	27.13	28.72	28.82	30.08	31.34	28.34	29.46	28.80	27.24	27.94	28.49	26.76	27.85

(LREE/HREE) = $\Sigma(\text{La-Sm})/\Sigma(\text{Gd-Lu})$; Subscript _{CN} refers to chondrite-normalised values; Subscript _{SN} refers to PAAS-normalised values.

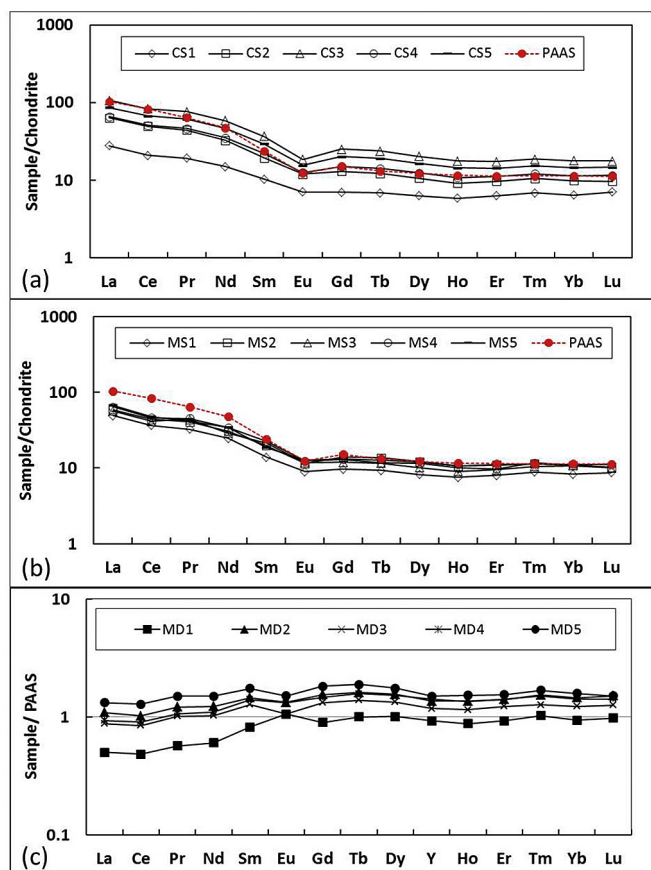


Fig. 7. Normalised REE patterns of the studied sediments. (a) Chondrite-normalised Ngwako Pan sandstones. (b). Chondrite-normalised D'Kar sandstones. (c) PAAS-normalised D'Kar Mudrocks.

6.2. Weathering intensity of the source area

The intensity of chemical weathering in the source area is controlled by the climate and the rate of tectonic uplift (Wronkiewicz and Condie, 1987). Therefore, information regarding the tectonic activity and climatic conditions in the source area can be obtained from weathering indices of sedimentary rocks. Moreover, modal compositions of sand can provide information on the intensity of chemical weathering as well as paleoclimatic conditions (Dickinson and Suczek, 1979; Van de Kamp, 2010). As already demonstrated in the Qm-F-Lt diagram (Fig. 3) the sandstones from both formations are classified as arkoses and subarkoses; with K-feldspar more abundant than plagioclase. According to Van de Kamp (2010), arkoses in which K-feldspar is dominant in abundance are formed in humid temperate and subtropical climates. It has to be noted that prolonged chemical weathering of feldspar-rich source rocks in the hot and humid environment will skew the modal composition of sandstones toward a quartz-bearing composition (e.g., Garzanti et al., 2013; Schneider et al., 2016).

As established by Nesbitt and Young (1982), the relationships among alkali and alkaline earth elements can be used as a measure of the intensity and duration of weathering. The mobility of these elements, especially during diagenesis can be evaluated by using the Chemical Index of Weathering (CIW; Harnois, 1988): $\text{Al}_2\text{O}_3/(\text{Al}_2\text{O}_3 + \text{CaO}^* + \text{Na}_2\text{O}) \times 100$ (in molecular proportions), where CaO^* is the CaO content within the silicate fraction only. The CaO detected in these rock samples is poorly correlated ($r = 0.23$, for the

sandstones of the Ngwako Pan Formation) to negatively correlated ($r = -0.91$, $r = -0.93$, for the sandstones and mudrocks of the D'Kar Formation, respectively) with Al_2O_3 implying that CaO is probably not incorporated in the silicate phases, but coming from calcite cements (see Section 4.1). Moreover, the investigated samples have highly variable CaO contents due to variation in calcite abundance and this may produce misleading conclusions if the CIW is used to infer the intensity of weathering (Cullers, 2000). To avoid the bias due to Ca distribution within the cements, Cullers (2000) proposed a modified Chemical Index of Weathering (CIW'), in which CaO is not considered: $\text{Al}_2\text{O}_3/(\text{Al}_2\text{O}_3 + \text{Na}_2\text{O}) \times 100$. The CIW' values of the investigated samples vary from 79 to 81 (average 79) for the sandstones of the Ngwako Pan Formation, from 77 to 80 (average 79) for the sandstones of the D'Kar Formation and from 81 to 89 (average 86) for the mudrocks of the D'Kar Formation. According to Cullers (2000) these high CIW' values are indicative of intense chemical weathering of the source rock, which may reflect a change of climate towards warm and humid conditions in the source area (Jacobson et al., 2003).

6.3. Provenance and source rock characteristics

Petrographic observations of the sandstones of both the Ngwako Pan and the D'Kar Formations point to a felsic igneous source rock with minor contributions from metamorphic or meta-sedimentary and sedimentary sources. A felsic volcanic source is supported by the high amount of monocrystalline inclusions-free quartz with straight extinction and the presence of volcanic rock fragments (Fig. 5a) in the sandstones of the D'Kar Formation (Blatt, 1992). The sandstones of the Ngwako Pan Formation, on the other hand, contain myrmekite and granitic clasts (Fig. 5b and c), medium to coarse-grained monocrystalline quartz containing microfractures-controlled fluid inclusions and exhibiting both undulose and non-undulose extinction, polycrystalline quartz composed of fewer crystals of more equant shape with little or no intercrystalline suturing (Fig. 4a). These characteristics, together with the presence of micropertite, point to a possible plutonic igneous source (Blatt, 1992). Furthermore, recrystallised quartz (Fig. 4d), polycrystalline quartz grains having polygonal crystal boundaries at 120° angle (Fig. 4c), and the presence of metamorphic rock fragments (Fig. 5d) in these sandstones reflect a substantial contribution from a metamorphic source rock. High amount of mudstone clasts (Fig. 5e) and the presence of microcrystalline chert grains (Fig. 5f), suggest a contribution of a proximal sedimentary source.

The felsic igneous provenance is further supported by major oxides and their ratios. TiO_2 -Zr plot (Fig. 9) shows that the source material for the sandstones of both formations is of felsic origin, whereas for the mudrocks it is of intermediate origin (Schreiber et al., 1992). In addition, many studies in tropical regions have shown that Al and Ti do not fractionate during sedimentary processes and the $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratios in the sediments can be considered as a pristine geochemical signature of the source rock composition (e.g., McLennan et al., 1983; Taylor and McLennan, 1985; Schieber, 1992; Garcia et al., 1994; Sugitani et al., 1996; Huntsman-Mapila et al., 2005; Paikaray et al., 2008). $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio normally varies from 3 to 8 in mafic rocks, 8–21 in intermediate rocks and 21–70 in felsic rocks (Sugitani et al., 1996, 2006; Hayashi et al., 1997). The $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio of the investigated samples shows values of 24.46–37.05, 21.36–29.29 and 18.55–21.32 for the sandstones of the Ngwako Pan Formation, the D'Kar Formation and mudrocks of the D'Kar Formation, respectively. The $\text{Al}_2\text{O}_3/\text{TiO}_2$ ratio indicates that the sandstones of both the Ngwako Pan and the D'Kar Formations were likely derived from felsic igneous rocks, whereas the chemical composition of the mudrocks of the D'Kar Formation is consistent with the contribution of an intermediate source rock.

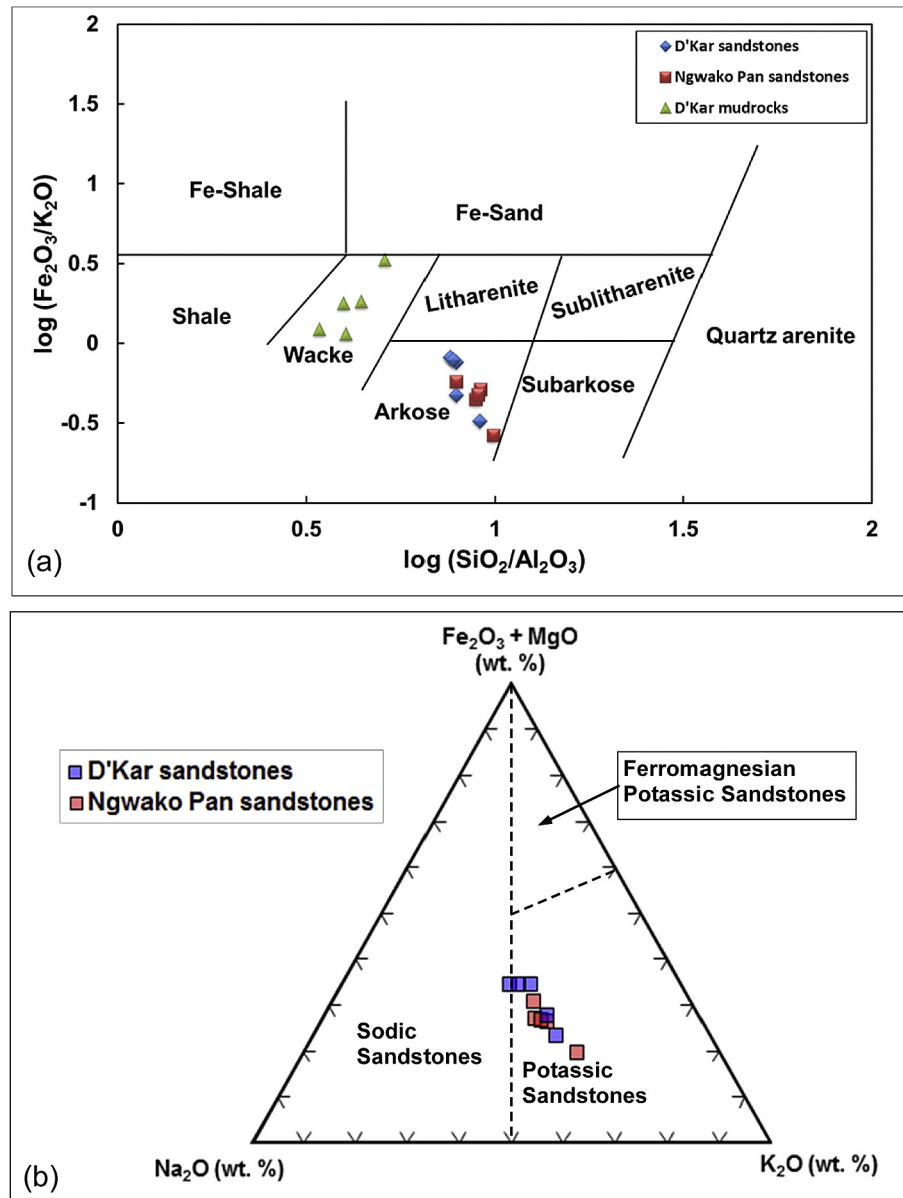


Fig. 8. Chemical classification of the investigated sandstones based on the [Herron \(1988\)](#) diagram. (b) Na_2O - K_2O - $(\text{Fe}_2\text{O}_3 + \text{MgO})$ ternary diagram showing the chemical composition of the sandstones from the D'Kar and Ngwako Formations (after [Blatt et al., 1972](#)).

The relative abundances of Cr, Co, Ni and V in siliciclastic sediments are useful in distinguishing mafic from felsic sources, as these elements are preferentially concentrated in mafic than felsic rocks ([Cullers, 1995](#); [Huntsman-Mapila et al., 2005](#); [Nagarajan et al., 2007](#); [Armstrong-Altrin et al., 2013, 2014, 2015](#)). The sandstones of both the Ngwako Pan and the D'Kar Formations have low concentrations of Cr, Co, Ni and V, demonstrating a felsic provenance. This is further supported by a significant positive correlation between Cr and Ni ($r = 0.99$, $n = 5$) demonstrated in D'Kar Formation ([Wrafter and Graham, 1989](#)). On the other hand, the mudrocks of the D'Kar Formation have higher concentrations of Cr, Co, Ni and V than the sandstones of both the Ngwako Pan and the D'Kar Formations ([Table 3](#)), but these values are lower than the sediments derived from mafic rocks. The Cr, Co, Ni and V values plus a negative correlation between Cr and Ni ($r = -0.63$, $n = 5$) demonstrated by the mudrocks from the D'Kar Formation suggest a mixed source. In

addition, the distribution of trace elements such as Ni and Cr may help further constraining the location of the source rocks. [Taylor and McLennan \(1985\)](#) suggested that Cr and Ni distributions such as those of Ngwako Pan and the D'Kar sandstones ([Fig. 10](#)) are consistent with a post-Archean source. This is further supported by ($\text{Gd}_\text{N}/\text{Yb}_\text{N}$) ratios that are less than 2 ([Table 4](#)), suggesting that they were derived from less HREE-depleted Archean and/or post-Archean sources ([Armstrong-Altrin et al., 2004](#)). Similarly, the La/Co , Th/Co , Th/Cr and Cr/Th ratios in felsic and mafic rocks show significant variations and as such are more useful in constraining the provenance of siliciclastic rocks ([Wronkiewicz and Condie, 1990](#); [Cox et al., 1995](#); [Cullers, 1995](#)). These ratios for the sandstones of both the Ngwako Pan and the D'Kar Formations are comparable to the range of sediments derived from felsic source rocks, whilst the mudrocks of the D'Kar Formation are comparable to sediments derived from mixed source rocks ([Table 5](#)).

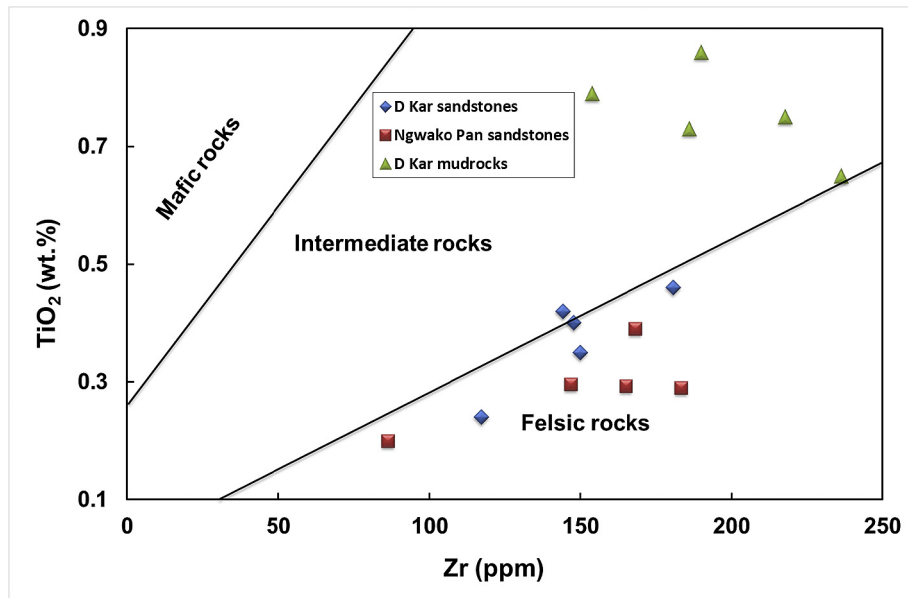


Fig. 9. TiO_2 vs Zr plot for the samples (after Hayashi et al., 1997).

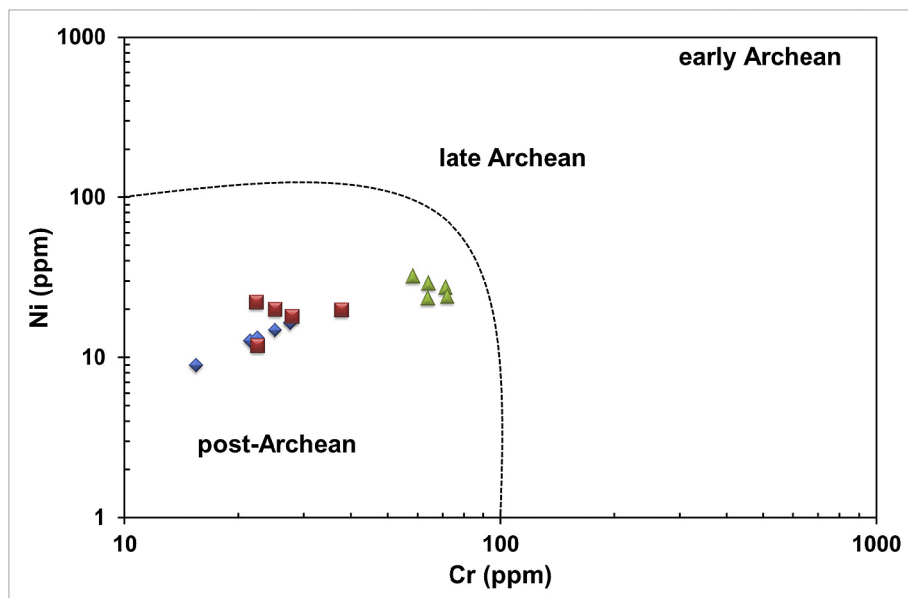


Fig. 10. Ni-Cr cross plot of major oxides and trace elements of D'Kar and Ngwako Pan sandstones and D'Kar mudrocks. Field boundaries after Taylor and McLennan (1985).

Table 5

Range of elemental ratios of the samples from this study compared to the ratios in similar fractions derived from felsic and mafic rocks.

Elemental ratio	Range of sediments		This study		
	Felsic Rocks ^a	Mafic Rocks ^a	Ngwako Pan sandstones	D'Kar sandstones	D'Kar mudrocks
Eu/Eu*	0.40–0.94	0.71–0.95	0.60–0.82	0.68–0.80	0.81–1.21
Th/Co	0.67–19.40	0.04–1.40	0.44–11.72	0.61–1.34	0.25–0.69
Th/Cr	0.13–2.70	0.02–0.05	0.18–0.52	0.17–0.30	0.09–0.18
Cr/Th	4.00–15.00	25.00–500.00	1.92–5.50	3.34–5.87	5.63–11.54

^a After Cullers and Podkovyrov, 2000.

Chondrite-normalised REE patterns and the size of the Eu anomaly have proven their utility as proxies for the assessment of the provenance of siliciclastic sediments (Taylor and McLennan,

1985; McLennan, 1989; Cullers and Podkovyrov, 2000; Singh, 2009; Shynu et al., 2011; Etemad-Saeed et al., 2011). Low LREE/HREE ratios and absence of Eu anomalies are characteristics of

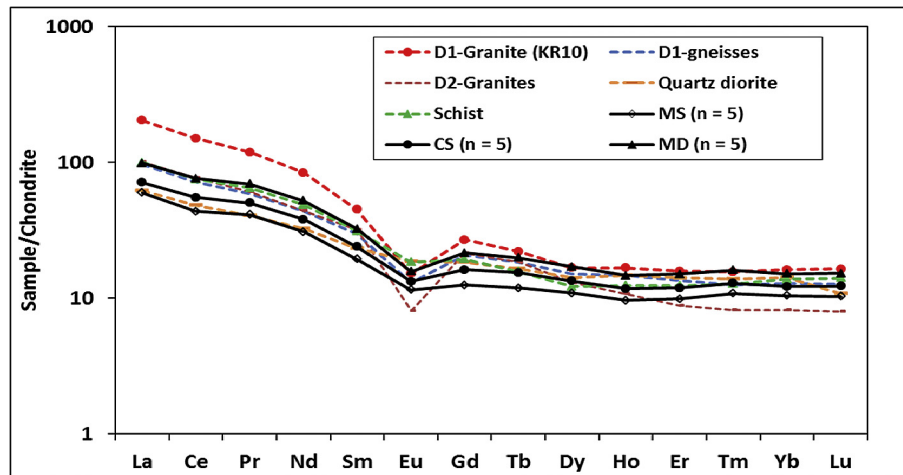


Fig. 11. Average chondrite-normalised REE patterns for Ngwako Pan and D'Kar sandstones, and D'Kar mudrocks compared to average chondrite-normalised REE patterns of pre-Neoproterozoic rocks (granites, gneisses, quartz diorite and schist) from different belts in Central Africa compiled from [Debruyne et al., 2015](#). Normalisation values are from [Taylor and McLennan, 1985](#). n = number of samples.

mafic rocks, whereas more felsic rocks are characterised by higher LREE/HREE ratios and negative Eu anomalies ([Cullers and Graf, 1984](#); [Cullers, 1994, 2000](#)). Furthermore, LREE enrichment, negative Eu anomaly and almost flat HREE patterns are indicative of cratonic source rock with the presence of felsic and meta-sedimentary components ([Gu et al., 2002](#); [Das et al., 2006](#)). Therefore, high LREE/HREE ratios ([Table 4](#)), significant negative Eu anomalies, LREE enrichment and almost flat HREE patterns ([Fig. 7a and b](#)) observed in the samples from both the Ngwako Pan and D'Kar Formations point to a source area of felsic composition. Moreover, chondritic Y/Ho ratios ([Table 4](#)) are typical for terrigenous materials derived from a magmatic source (e.g., [Klinkhammer et al., 1994](#); [Bau and Dulski, 1999](#)). Accordingly, the average Y/Ho values of the studied samples are consistently below 30 supporting their derivation from a magmatic source.

To further underpin the provenance identification, the average REE data for pre-Neoproterozoic rocks (granites, gneisses, quartz diorite and schist) from Mesoproterozoic Kibara Belt (KIB) and the Karagwe-Ankole Belt in Central Africa ([Debruyne et al., 2015](#)) were compared with the average REE data of the investigated samples ([Fig. 11](#)). The general shapes of the REE pattern of the investigated samples are comparable to the average D1-granite, D1-gneisses and D2-granites. This comparison suggests that the sediments for the investigated samples were derived largely from felsic rocks.

From the discussion elaborated above, both petrographic and geochemical data suggest that the sediments of the Ngwako Pan and the D'Kar Formations were largely derived from felsic igneous rocks with minor contribution from metamorphic and sedimentary rocks. Hence, considering the paleogeography of the area during the Neoproterozoic, the probable source rocks are located in the Palaeoproterozoic basement rocks (granitoids and granitic gneiss) and the Mesoproterozoic Kgwebe volcanic rocks exposed north from the study area ([Kampunzu et al., 1998](#); [Modie, 2000](#)).

6.4. Tectonic setting

The chemical and petrographic composition of the sandstone can also reflect the tectonic setting of the source rock during the weathering and erosion ([Dickinson, 1985](#)). The Qt-F-L and Qm-F-Lt

diagrams ([Fig. 12a and b](#)) indicate that the sandstones were probably derived from a continental block provenance ([Dickinson, 1985](#)). As indicated by [Blatt \(1992\)](#) and [Maynard \(1984\)](#), the enrichment of quartzo-feldspathic detritus coupled with high K-feldspar/plagioclase ratio in the sandstones is allusive of sands derived from rifted areas associated with continental separations.

During the past 25 years, major elements patterns of clastic rocks were used to discriminate the tectonic settings of the ancient siliciclastic depositional systems. The discrimination was based on the K_2O/Na_2O ratio and the SiO_2 content, as they vary consistently in different tectonic environments ([Roser and Korsch, 1986](#); [Maynard et al., 1982](#)). However, the use of these discrimination diagrams has recently received massive criticism from numerous workers (e.g., [Armstrong-Altrin and Verma, 2005](#); [Armstrong-Altrin, 2015](#); [Weltje, 2006](#); [Ryan and Williams, 2007](#); [von Eynatten et al., 2012](#)) as their performance is unsatisfactory. Therefore, in this study, new discriminant-function-based multi-dimensional diagrams recently introduced by [Verma and Armstrong-Altrin \(2013\)](#) were used to decipher the tectonic setting of the investigated samples. These diagrams which discriminate three tectonic settings, viz, arc (active volcanism), continental rift (extension) and continental collision (compression) were constructed from $\log_e$ -ratio transformation of all major elements. In these diagrams major elements (SiO_2 to P_2O_5) are adjusted to 100% on a volatile free basis, resulting in two multi-dimensional diagrams for high-silica ($(SiO_2)_{adj} = 63\text{--}95\%$) and low-silica ($(SiO_2)_{adj} = 35\text{--}63\%$) clastic sediments. Since the adjusted SiO_2 concentration is greater than 63%, all the investigated samples from both formations plots in the high-silica diagram ([Fig. 12c](#)). In this diagram, the majority of the sandstones of both the Ngwako Pan and the D'Kar Formations plot in the field for a rift setting suggesting a passive margin ([Verma and Armstrong-Altrin, 2016](#)), except for two samples (one sample each) from both the Ngwako Pan and the D'Kar Formations that lie in the collision field, close to the boundary between the collision and the rift field. This result is consistent with the result inferred from the Q-F-L ternary diagrams ([Dickinson, 1985](#)) in which the sandstones of both the Ngwako Pan and the D'Kar Formations plot in the continental block (transitional continental) field ([Fig. 12a, b](#)). Moreover, a continental rift setting inferred from the high silica diagram is consistent with the geology of the area ([Modie, 1996](#)).

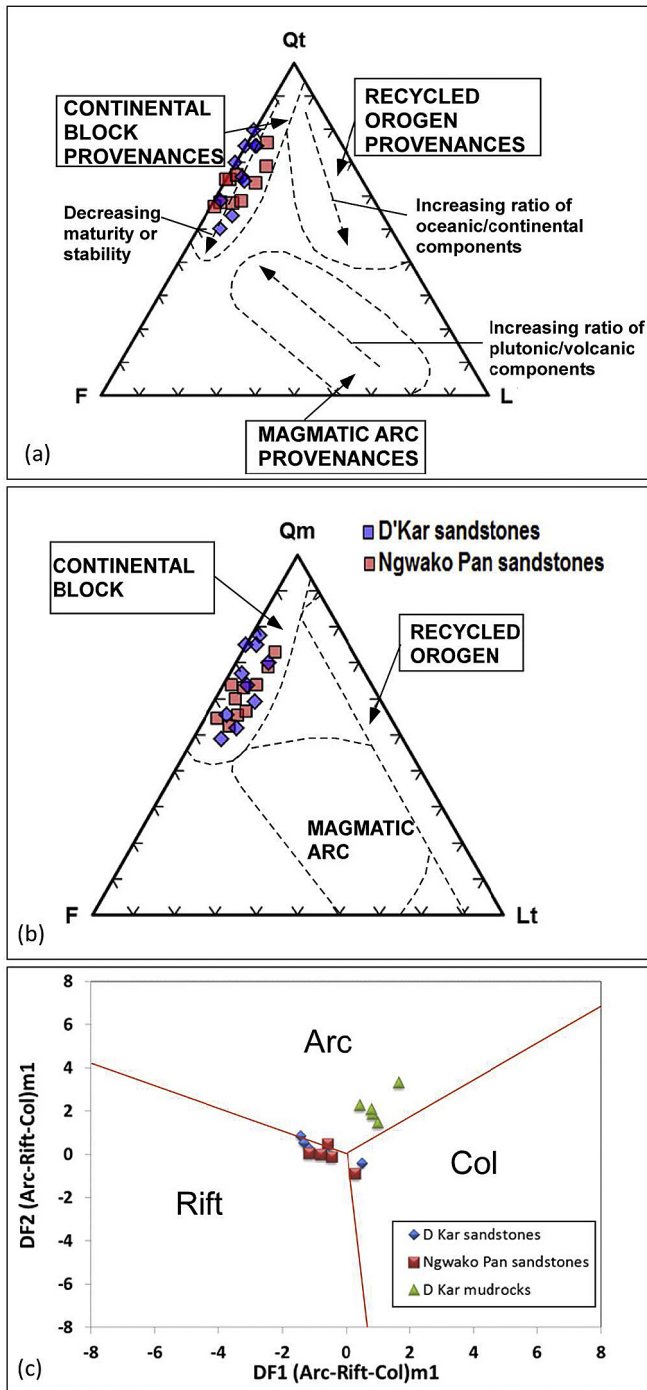


Fig. 12. (a, b) Qt-F-L and Qm-F-Lt ternary plots for palaeotectonic provenance of the studied sandstones (after Dickinson, 1985). Qt = Total quartz grains (monocrystalline + polycrystalline), F = Total feldspar grains (plagioclase + K-feldspar), L = Unstable siliciclastic lithic fragments. (c) Discriminant function multi-dimensional diagram for high-silica clastic sediments (Verma and Armstrong-Altrin, 2013). The discriminant functions DF1 and DF2 equations for high-silica clastic sediments are: $DF1_{(Arc-Rift-Col)m1} = (-0.263 \times \ln(TiO_2/SiO_2)_{adj}) + (0.604 \times \ln(Al_2O_3/SiO_2)_{adj}) + (-1.725 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (0.660 \times \ln(MnO/SiO_2)_{adj}) + (2.191 \times \ln(MgO/SiO_2)_{adj}) + (0.144 \times \ln(CaO/SiO_2)_{adj}) + (-1.304 \times \ln(Na_2O/SiO_2)_{adj}) + (0.054 \times \ln(K_2O/SiO_2)_{adj}) + (-0.330 \times \ln(P_2O_5/SiO_2)_{adj}) + 1.588$; $DF2_{(Arc-Rift-Col)m1} = (-1.196 \times \ln(TiO_2/SiO_2)_{adj}) + (1.064 \times \ln(Al_2O_3/SiO_2)_{adj}) + (0.303 \times \ln(Fe_2O_3t/SiO_2)_{adj}) + (0.436 \times \ln(MnO/SiO_2)_{adj}) + (0.838 \times \ln(MgO/SiO_2)_{adj}) + (-0.407 \times \ln(CaO/SiO_2)_{adj}) + (1.021 \times \ln(Na_2O/SiO_2)_{adj}) + (-1.706 \times \ln(K_2O/SiO_2)_{adj}) + (-0.126 \times \ln(P_2O_5/SiO_2)_{adj}) - 1.068$.

On the other hand, the mudrocks of the D'Kar Formation plot in the field for arc setting suggesting an active margin (Verma and Armstrong-Altrin, 2016).

7. Conclusive remarks

Here we presented a full array of petrographic and geochemical data aiming at the characterising the Ngwako Pan and D'Kar Formations (Ghanzi Group) in north-west Botswana. These siliciclastic sediments host the Banana Zone Cu-Ag mineralisation, which is part of the Kalahari Copperbelt of northwest Botswana.

Based on the petrographic and geochemical characteristics of the studied sandstones and mudrocks, the following conclusions can be drawn:

1. Sandstones of both the Ngwako Pan and D'Kar Formations are potassic and classified as subarkoses, and arkoses. These sandstones are generally considered as chemically and texturally immature to submature.
2. The major oxides and trace elements cross-plots as well as the REE normalised patterns demonstrate that the detritus were largely from felsic igneous source rocks with subordinate sedimentary or meta-sedimentary, intermediate volcanic and metamorphic rocks.
3. The mudrocks of the D'Kar Formation are likely derived from intermediate igneous rocks.
4. The relatively high modified Chemical Index of Weathering yielded by the studied sandstones (79–81, 77–80 and 81–86 for the sandstones of the Ngwako Pan Formation, the sandstones of the D'kar Formation and mudrocks of the D'kar Formation, respectively) suggest that the terrigenous materials were supplied from a highly weathered continental block.
5. As suggested by the high-silica diagram, the siliciclastic system was probably settled in a rift environment (passive margin).
6. The probable source rocks of the sandstones of both the Ngwako Pan and the D'Kar Formations are located in the Palaeoproterozoic basement rocks (granitoids and granitic gneiss) and the Mesoproterozoic Kgwebe volcanic rocks exposed north of the study area.

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