GEOCHRONOLOGY, MAGNETIC LITHOSTRATIGRAPHY, AND THE TECTONOSTRATIGRAPHIC EVOLUTION OF THE LATE MESO-TO NEOPROTEROZOIC GHANZI BASIN IN BOTSWANA AND NAMIBIA, AND IMPLICATIONS FOR COPPER-SILVER MINERALIZATION IN THE KALAHARI COPPERBELT

By

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ABSTRACT

Despite a wealth of research on the Kalahari Copperbelt over the past 30 years, two crucial aspects of the mineralizing systems have remained elusive. First, the age of the rift sequence hosting the deposits and, second, the nature of the fluid pathways for the mineralizing fluids.

Laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS) U-Pb isotopic analysis on one igneous sample of the Makgabana Hills rhyolite (Kgwebe Formation) within the central Kalahari Copperbelt in Botswana constrains the depositional age of the unconformably overlying Ghanzi Group to after 1085.5 ± 4.5 Ma. The statistically youngest detrital zircon age populations obtained from the uppermost unit of the Ngwako Pan Formation (1066 ± 9.4 Ma, MSWD = 0.88, n = 3), the overlying D’Kar Formation (1063 ± 11, MSWD = 0.056, n = 3), and the lower Mamuno Formation (1056.0 ± 9.9 Ma, MSWD = 0.68, n = 4) indicate that the middle and upper Ghanzi Groups were deposited after ~1060 to ~1050 Ma. Lu-Hf analysis of detrital zircon suggests that the Mesoproterozoic and Paleoproterozoic rocks of the Namaqua Sector and the Rehoboth Basement Inlier were the primary sediment sources for the siliciclastic rocks of the Ghanzi Group and lesser material was derived from the basin-bounding footwall margin of the northwest Botswana rift, the Paleoproterozoic Magondi Belt and the Okwa Block, and possibly parts of the Limpopo Belt on the northern margin of the Kalahari Craton in southern Africa.

A molybdenite Re-Os age of 981 ± 3 Ma provides a minimum depositional age constraint on D’Kar Formation sedimentation. Authigenic xenotime U-Th-Pb ages of ~925 and 950 Ma further the evidence for an earliest Neoproterozoic (Tonian) age for the D’Kar Formation. Re-Os ages of 549 ± 11.2 Ma (low-level highly radiogenic chalcocite-idaite) and 515.9 ± 2.4 Ma (molybdenite), and a U-Th-Pb age of 538.4 ± 8.3 Ma (xenotime inclusion in chalcopyrite) from several Cu-Ag deposits in the central Kalahari Copperbelt suggest prolonged mineralizing events during basin inversion related to the Pan-African (~600 to 480 Ma) Damara orogen.

High-resolution aeromagnetic maps were utilized to define the stratigraphy and structure of the Ghanzi Group of the central Kalahari Copperbelt. Maps of the second vertical derivative transformation were compared with detailed stratigraphic data from drill holes. These data reveal previously unrecognized thinning of the Ngwako Pan Formation onto rocks of the underlying Kgwebe Formation and suggest the presence of syn-sedimentary horst and graben and/or half-graben structures. Truncation of the aeromagnetics fabric of the uppermost Ngwako Pan Formation rocks suggests that the rocks were gently folded and eroded above palaeotopographic highs prior to the ensuing basin-wide marine transgression and sedimentation of the unconformably overlying mixed marine siliciclastic and carbonate rocks of the D’Kar Formation. Detailed facies architecture derived from both drilling and magnetic data at the Zone 5 Cu-Ag deposit, located east of the Kgwebe and Makgabana Hills, suggests that its physical
(stratigraphic) and chemical (organic material and \textit{in-situ} bacteriogenic sulfide) nature were influenced by the underlying basin architecture, which was critical in development of trap sites and in funneling epigenetic hydrothermal fluids into those traps.

The presented new data indicate that the basin architecture underlying the Kalahari Copperbelt strongly influenced the localization of deformation and hydrothermal fluid flow during epigenetic events. The results of this study can be used to help vector exploration for Cu-Ag deposits through the recognition of buried paleotopographic highs and associated favorable trap sites.
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CHAPTER 1

GENERAL INTRODUCTION

1.1 Introduction

The Kalahari Basin (Kalahari Desert) in southern Africa stretches approximately 2200 km from South Africa in the south, northwards through Botswana and Namibia, and up into Angola. Unconsolidated sand of the Kalahari Group covers some 75% of Botswana (Figure 1.1, Haddon, 2005). The Precambrian Kalahari Copperbelt metallogenic province in northwestern Botswana and central Namibia is predominantly concealed by this Kalahari Group cover.

The Kalahari Copperbelt contains several sedimentary rock-hosted copper-silver deposits and prospects that have many similarities to those in the world class mining districts of the European Kupferschiefer and Central African Copperbelt. Deposits in the belt range in size from ~10 Mt to ~100 Mt with grades of ~1.0 to ~2.4% copper and 10 to 40 grams per ton silver. To date, total mineral resources in the Ghanzi Ridge area of northwestern Botswana total more than 6.5 million tons of contained copper (measured, indicated, and inferred mineral resources as of 2017, Cupric Canyon Capital, https://www.cupriccanyon.com/development-exploration/exploration).

The extensive Kalahari Group cover has limited efforts to conduct detailed regional-scale studies on the rocks hosting the copper deposits. Geophysical surveys including regional fixed-wing, district-scale high-resolution helicopter-borne, and ground magnetics as well as gravity surveys have been conducted over the region in recent years by Botswana government and mineral exploration companies to better understand the regional geology under cover. Regional exploration and recent mine development within the Lake N’Gami sector of the Ghanzi Ridge has produced an abundance of geologic data that has been utilized to better understand the relationships between the geology and metallogeny of the Kalahari Copperbelt on multiple scales. Several aspects of the mineralizing systems have been studied in detail over the past 30 years from drill cores including metal and sulfur sources, the nature of the mineralizing fluids, the drivers behind hydrothermal fluid flow, and the physical and chemical processes to trap mineralizing fluids. The previous research has led to a more comprehensive understanding of the deposit-scale controls on the copper-silver deposits. However, two critical aspects of the mineralizing systems have remained elusive: a well-constrained age for the host rocks and the pathways that channeled mineralizing fluids to trap sites.

The Meso- to Neoproterozoic Ghanzi Group metasedimentary rocks described in this study occur in the Lake N’Gami area, or northeastern sector of the Ghanzi Ridge in northwestern Botswana (Figure 1.1). The Ghanzi Ridge is a discontinuous, ~450-km-long, 30- to 60-km wide, northeast-southwest
Figure 1.1 The Kalahari Basin and the Kalahari Copperbelt. Left: locality map of the Kalahari Basin in southern Africa annotated with the location of the Kalahari Copperbelt in Namibia and Botswana. Modified from Haddon (2005). Right: Satellite image displaying the geophysically defined limits of the Kalahari Copperbelt (red), shallowly covered basement rocks of the Ghanzi Ridge (yellow line) and its sub-sectors (dashed white lines), and locations of Cu-Ag mines, resources, and prospects. This research focuses on the Lake N’Gami (northeastern) sector of the Ghanzi Ridge (white box).
trending structural window of deformed and peneplained Precambrian rocks that are overlain by shallow (<1 meter to ~40-meters-thick) Miocene-present Kalahari Group aeolian sand within the south-central Kalahari Basin (Figure 1.1).

The deformed Ghanzi Group rocks occur in the southwestern half of the Ghanzi-Chobe zone, a 700-km-long and 150- to 200-km-wide, northeast-southwest-trending belt of geophysically defined deformed basement rocks that extends from the Chobe district in northern Botswana to the western border with Namibia in the Ghanzi and Ngamiland districts (Figure 1.2; Reeves and Hutchins, 1982; Meixner, 1983; Modie, 1996). The metasedimentary host rocks were folded together with underlying metavolcanic rocks, some of which form isolated chains of basement inliers in the Lake N’Gami area of the Ghanzi Ridge.

The Chobe and Ghanzi/Lake N’Gami volcano-sedimentary basins in northern and western Botswana, respectively (Borg, 1988), are collectively referred to as the northwest Botswana rift (Key and Ayres, 2000). The northwest Botswana rift formed the northeastern extension of a series of inverted Precambrian volcano-sedimentary rift basins that extend for another ~400 kilometers into central Namibia south of the town of Rehoboth (Figure 1.2; Borg, 1988; Lehmann et al., 2015). There, the fold belt is referred to as the Southern Margin zone of the Damara Belt, a Pan-African (~600 to ~480 Ma) tectonic zone formed during the suturing of the Kalahari and Congo cratons as Gondwana was assembled (Toens, 1975; Martin and Porada, 1977a, b; Miller, 1983, 2008).

Inversion of rift basins along the leading margins of both these cratons during continental assembly is regarded as a trigger for widespread hydrothermal base and precious metal mineralizing events within the basins (e.g. Sillitoe et al., 2010; Maiden and Borg, 2011). In districts around the world, the locations of many sedimentary rock-hosted copper deposits have been related to syn-sedimentary basin-architecture faults. Such faults have been inferred to be responsible for the development of sedimentological and/or chemical trap sites as well as acting as fluid conduits that focused oxidized, metal-bearing fluids during multiple stages of basin development spanning early to late diagenesis and basin inversion (Hitzman et al., 2005).

The objectives of this study are two-fold: first, to gain a better understanding of the tectono-stratigraphic evolution of the Ghanzi basin in the Ghanzi Ridge area with emphasis on determining the age of sediment deposition and timing of mineralizing events as well as characterizing the stratigraphy and structure of the basin, and second, to explore the possible role that basin architecture may have played in the metallogeny of the Kalahari Copperbelt. A multi-disciplinary approach was employed that utilized available geologic data from mineral exploration companies to:

- assess the deposit- to regional-scale sedimentology and stratigraphy of the host rocks,
Figure 1.2 Location map of the primary study area (red box) within the Precambrian crustal framework of southern Africa. The Kalahari Copperbelt occurs within the Ghanzi-Chobe zone (GCz) and Southern Foreland zone (SFz) of the Damara orogeny, which is part of the broader Pan-African (~600 to ~480 Ma) mobile belt system. Dark gray indicates areas of shallowly covered to exposed basement rocks in the Kalahari Copperbelt. Abbreviations: CACB = Central African Copperbelt; CKB = Choma-Kalomo Block; ZB = Zambezi Belt. Modified from Borg (1988), Key and Ayres (2000), Jacobs et al. (2008); and Lehmann et al. (2015).

- conduct U-Pb zircon geochronology on the meta-volcanic rocks below the host rock sequence to provide a maximum depositional age for the host rocks
- perform U-Pb and Lu-Hf zircon geochronology of detrital zircons from the meta-sedimentary host rocks to place age constraints on the timing of different stages of basin development and interpret provenance sources for the metasedimentary host rocks,
- undertake Re-Os and U-Th-Pb radiogenic and sulfur stable isotopic studies to place age constraints of the timing of mineralizing events in the basin, and
- process, model, and interpret multi-scale geophysical data to aid in deposit- to regional-scale geologic mapping of the host rocks and structural elements of the Ghanzi Ridge area,
Practically, the purpose of the research was to provide a better regional context to highlight areas with mineral potential, particularly in portions of the belt where geochemical datasets and shallow drilling are ineffective (i.e. mineralized horizons concealed by overlying geologic formations).

1.2 Dissertation organization

This dissertation is divided into three parts: an introduction and literature review that outlines the objectives of the study with a brief background on the geologic framework of the study area and the methods utilized during the research, a main body composed of three original research papers conducted by the author, the author’s advisors and committee members, and various collaborators, and a discussion of the overall results and implications of the research.

1.2.1 Igneous and detrital zircon geochronology

To help resolve the question of the timing of basin development in the Ghanzi Ridge area, both igneous and sedimentary rock samples were collected from the Ghanzi basin for zircon U-Pb analysis. Separation of the zircon from the igneous and sedimentary rock samples was performed by co-author Dr. Richard Moscati of the United States Geological Survey in Denver. U-Pb analyses of igneous zircon were undertaken at the United States Geological Survey, Denver, by co-author Dr. Christopher Holm-Denoma. U-Pb analyses were carried out at the University of California Santa Barbara by Dr. Andrew Kylander-Clark, who co-authored the paper on zircon geochronology.

LA-ICPMS U-Pb analyses allow for rapid analysis of many detrital zircon grains (Gehrels, 2012). U-Pb zircon geochronology was applied to the selected samples of the Ghanzi Group to:

- determine the maximum depositional age, where the youngest age component in a clastic unit provides the earliest possible age of deposition, and
- conduct a provenance study, where ages of detrital minerals are compared with ages of potential source terranes to determine sources of sediment

To aid in the provenance study, Lu-Hf geochronology was employed on the detrital zircon suite. Lu-Hf isotope systematics in zircon can be used to determine whether the magmas that generated the zircon in the source region were sourced from juvenile or reworked crust. When combined with detrital zircon U-Pb ages, the mode of crustal formation (juvenile vs. reworked) in each magmatic generation and the continental growth rate in the source region can be estimated (Iizuka et al., 2005). The robustness of zircon to later metamorphism gives the Lu-Hf method an advantage over whole-rock isotopic systems such as Sm-Nd or Rb-Sr geochronology. Hf model ages have been shown to be in good agreement with undisturbed whole-rock Nd model ages (Iizuka et al., 2005). Detrital zircon U-Pb and Lu-Hf data can be a powerful tool when compared against databases of magmatic isotopic data and model ages. Both U-Pb and Lu-Hf data from the Ghanzi Group were compared to an exhaustive compilation of U-Pb, Lu-Hf, and
Sm-Nd age data from southern African terranes. The combined datasets were then utilized to interpret the
tectono-stratigraphic development of the basin over time by looking at aspects such as differing
provenance(s) for successive formation/rock units, primary and secondary sediment sources, sediment
transport direction, and sedimentary architecture (i.e. proximal versus distal parts of the basin).

1.2.2 Xenotime U-Th-Pb geochronology

Diagenetic phosphate minerals are widespread in siliciclastic sedimentary rocks of all ages and
form during various stages of basin evolution (Fletcher et al. 2000). Monazite and xenotime are ideal
minerals for U-Th-Pb isotopic dating because of its relatively high U and Th contents, but typically low
concentration of common Pb (Liu et al., 2005). Small, early diagenetic xenotime overgrowths commonly
occur on zircons in siliciclastic sedimentary rocks. U-Pb dating of xenotime has been proven to be
particularly useful in Precambrian terranes, where diagenetic xenotime has been used as an equivalent to
biostratigraphic dating in the Phanerozoic.

Electron microprobe analyses of xenotime grains were undertaken by Dr. Heather Lowers of the
United States Geological Survey to ensure that the correct spots identified by electron dispersive
spectrometry were analyzed by LA-ICPMS. U-Pb analyses were carried out at the University of
California Santa Barbara by Dr. Andrew Kylander-Clark, who is included as a co-author. LA-ICPMS U-
Th-Pb geochronology was carried out on xenotime grains identified using scanning electron microscopy
(SEM) and elemental dispersive spectrometry (EDS) analyses in order to determine constraint the ages of
deposition of sedimentary rocks of the Ghanzi basin and diagenetic and epigenetic processes that affected
them.

1.2.3 Sulfide Re-Os geochronology

Re and Os generally occur at the ppt level in the earth’s crust with three notable exceptions –
organic-rich mudstones, sulfides, and heavy hydrocarbons. Rhenium-osmium (Re-Os) geochronometry
is the first widely applicable method for direct dating of ore minerals (certain sulfides and oxides),
particularly molybdenite, which is the only naturally occurring mineral in Earth’s crust that consistently
takes in Re at the ppm level while essentially excluding Os from its structure upon formation (Stein,
2014). The lack of initial or common Os coupled with ppm levels of Re (Re/Os > 10⁶) yields readily
measurable radiogenic ¹⁸⁷Os in a geologic sample (Stein et al., 1997, 2001). The Re-Os chronometer has
proven to be robust in most geological situations, as precise isochrons have been obtained even for
molybdenites that have undergone high-grade metamorphism and deformation (Raith and Stein, 2000;
Bingen and Stein, 2003). This is because Os does not leave a crystal during metamorphism as it is highly
insoluble in most minerals or reduced aqueous solutions (Stein, 2014).
Low level, highly radiogenic (LLHR) sulfides are like molybdenite in that they may contain high Re/Os ratios ($^{187}$Re/$^{188}$Os in the thousands) and minimal common or initial Os upon crystallization (Stein, 2014). LLHR pyrite and arsenopyrite are particularly useful because these sulfides are common minerals in many ore deposits. However, cases where both parent and daughter elements are incorporated into a sulfide on crystallization requires the isochron approach to acquire a model age.

More recently, Re-Os geochronology has been utilized to determine the timing of hydrocarbon maturation (oil, bitumen) and unravel important information about organic-rich source rocks and maturation-migration histories of hydrocarbons within sedimentary basins (Finlay et al., 2011, 2012; Stein and Hannah, 2014). These developments have sparked interest into the relationships between hydrocarbons and base metal deposits in sedimentary basins; recent research suggests that the fluids responsible for many sediment-hosted metallic ore deposits may have a direct link to multiple phases of production and migration of hydrocarbons (e.g. Holdsworth et al., 2013). Proterozoic basins present a more difficult area of study because much of the hydrocarbons generated in such basins may have been expelled or converted to graphite during subsequent metamorphism and deformation. However, sulfides with hydrocarbons within these basins could act as tracers of tectonic and thermal events that affected the basin, including hydrocarbon production and migration.

Re-Os geochronology was carried out by co-author Dr. Holly Stein and the staff of the A.I.R.I.E. Program at Colorado State University. Re-Os geochronology was conducted on molybdenite, copper sulfides (chalcopyrite, bornite, chalcopyrite), and arsenopyrite to build upon previous Re-Os results (e.g. Hall, 2013) and helped refine the timing of mineralizing events. Evidence for the proposed remobilization of diagenetic sulfides was studied though a combination of petrographic methods to characterize the mineralogy and textural characteristics of the sulfide minerals. Sulfur stable isotopic analysis was undertaken at the United States Geological Survey, Denver, by Dr. Cayce Gulbransen and Dr. Craig Johnson. Sulfur isotopic analysis was employed to determine the sulfur source for the dated sulfides. These data helped refine the timing of plausible diagenetic and epigenetic events in Ghanzi basin and how they related to the overall tectonic configurations of the Kalahari Craton through time.

1.2.4 Geophysics - magnetics

Since 2010, mineral exploration companies have used high-resolution helicopter-borne and ground magnetics datasets to delineate the prospective Ngwako Pan – D’Kar Formation contact with great accuracy (within 20-30 meters) under cover of >40 meters. Various traditional filters applied to the Total Magnetic Intensity (TMI) datasets included reduction-to-pole (RTP), the analytic signal (AS), first vertical derivative (1VD), and tilt derivatives (TDR). The filtered aeromagnetic maps either accentuate or depress long- and short-wavelength anomalies, which can then be interpreted based on known geology.
The numerically transformations aided in the production of detailed aeromagnetic maps that resolved not only the contacts between the main stratigraphic formations, but also the intraformational magnetic fabrics of the different rock formations. Magnetic lineaments within these datasets also aided in identifying structural elements such as the position of fold axes and previously unknown faults (Discovery Metals Limited internal reports, 2010-2012). Qualitative observations from well-constrained geology and an aeromagnetic survey over the Boseto copper deposits suggest that the second vertical derivative magnetic fabric of the D’Kar Formation coincided well with the known stratigraphy (Hall and Hitzman, 2016). The findings suggest that the aeromagnetic datasets could be used to produce detailed geologic maps of the Ghanzi-Chobe zone that can be used to further interpret the geology under cover.

The observed magnetic fabrics was compared with well-constrained stratigraphy from drill holes at several deposits and districts in the Ghanzi Ridge area. The two datasets were first compared qualitatively to establish relationships between lithostratigraphy and magnetic fabrics. Qualitative relationships were then tested quantitatively through collection of petrophysical data from borehole samples. The petrophysical data was modelled to determine what, if any, characteristics of the rock magnetization are responsible for the observed relationships. The ultimate aim was to produce highly detailed stratigraphic and structural maps across the basin to enable interpretations of basin architecture and its evolution through time and the determine any spatial relationships between stratigraphic architecture, structure, and the locations of known ore deposits.

1.2.5 General conclusions

The combined geochronology and aeromagnetic interpretations were evaluated in terms of the tectonostratigraphic evolution of the Ghanzi basin. Age constraints on the different stages of basin evolution in the northwest Botswana rift are discussed based on the results of geochronological investigations. The influence of the underlying basin architecture on the development of trap sites and fluid pathways is highlighted. The age of mineralizing events is discussed in terms of the evolution of the basin from initial formation to inversion.

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CHAPTER 2
LITERATURE REVIEW

2.1 Introduction

The Kalahari Copperbelt in Botswana is poorly studied due to widespread cover of the Kalahari Desert. Thus, literature on the geology is restricted to a few published papers regarding the regional geology and mineralization. Much of the known geology in the Ghanzi Ridge study area has come from mineral exploration company geological reports and graduate student research projects completed in recent years. This chapter reviews the available literature on regional- to deposit-scale geology with a focus on the topics to be covered in this dissertation. Characteristics and genetic models for the sedimentary rock-hosted stratiform copper ore system was reviewed first to better understand the context of sedimentary rock-hosted stratiform copper deposits and lay the groundwork for the topics to be addressed by this study – understanding basin architecture, basin evolution, and the timing of diagenetic to epigenetic mineralizing events. This is followed by a discussion of the regional geologic framework of the Kalahari Copperbelt, previous geophysical exploration and the exploration and development history of copper deposits, and finally the characteristics of the sedimentary rock-hosted copper deposits in Botswana.

2.2 The sedimentary rock-hosted stratiform copper ore system

Sediment-hosted stratiform copper deposits account for approximately 23 percent of the world’s Cu production and known reserves in addition to being significant sources of Co and Ag. (Hitzman et al., 2005). The copper deposits comprise relatively thin (generally <30 m and commonly less than 3 m), disseminated to veinlet Cu and Cu-Fe sulfide-bearing zones that are peneconcordant with lithological layering in siliciclastic or dolomitic sedimentary rocks.

Deposits are the products of evolving basin-scale fluid-flow systems that include source(s) of metal and S, source(s) of metal- and S-transporting fluids, the transport paths of these fluids, a thermal and/or hydraulic pump to collect and drive the fluids, and the chemical and physical processes which result in precipitation of the sulfides. Metal sources are undoubtedly red-bed sedimentary rocks containing Fe oxyhydroxides capable of weakly binding metals (Figure 2.1). Sulfur may be derived from marine or lacustrine evaporites, reduced seawater, or hydrogen sulfide-bearing petroleum. Metals are interpreted to have been transported at low to moderate temperatures in moderately to highly saline aqueous fluids, with the temperature of the fluid largely dependent on the time of fluid migration in the basin’s burial history. These basinal fluids were focused to potential metal precipitation sites by thinning
Figure 2.1 Schematic cross section across an intracratonic, hydrologically closed basin that is typical of those hosting sediment-hosted stratiform copper deposits. The diagram highlights the various geologic factors and processes involved in mineralizing system. From Hitzman et al. (2010).

of the red-bed sequence at basin margins, by faults, by differentially permeable sedimentary units, by paleotopography within the basin, or along the margins of salt diapirs (Figure 2.2).

Sulfide precipitation occurred due to reduction, typically caused by reaction with carbonaceous rocks or petroleum. The amount of sulfides present at any deposit may be either metals or sulfide limited or could have been controlled by the amount of available reductant. While the understanding of sediment-hosted stratiform copper ore genesis at the deposit scale is relatively robust, there are still significant questions in regards its position in terms of basin evolution. A wide variety of basin architectures and processes can lead to the formation of sediment-hosted stratiform copper deposits (Hitzman et al., 2005). Despite general agreement that sulfides postdate sedimentation, the absolute age of mineralization in many deposits has been difficult to document and the available evidence suggests that deposits can form throughout a basin’s evolution from early diagenesis of ore host sediments to basin inversion and metamorphism (Hitzman et al., 2010). Periods in Earth’s history of supercontinent break-up favored the formation of failed rifts that subsequently became significant intracratonic basins with basal, synrift red-bed sequences overlain by marine and/or lacustrine sediments and, in some basins located at low latitudes, by thick evaporitic strata (Figure 2.3). The intracratonic setting of these basins allowed the development of a hydrologically closed basinal architecture in which highly oxidized and saline, moderate-temperature basinal brines were produced that were capable of supplying reduction-controlled
Figure 2.2 Cross-sectional models of basin-scale fluid flow in sediment-hosted stratiform copper systems. 
sulfide precipitation over very long periods of time (tens to hundreds of millions of years). The length of time available for the mineralizing process to operate may be the key factor in forming supergiant deposits. Glaciation may also be conducive for the formation of supergiant sediment-hosted stratiform copper deposits. Glacial periods correspond to magnesium- and sulfate-rich oceans that could have been responsible for additional sulfur in basinal brines developed during evaporite formation and would then be available during the long mineralization process (Hitzman et al., 2010).

2.3 Geologic framework of the Kalahari Copperbelt in Botswana

Sedimentary rock-hosted copper deposits of the Kalahari Copperbelt in Botswana occur within the Ghanzi-Chobe zone, a 600-km-long by 150- to 200-km-wide fold belt situated on the northwestern margin of the Kalahari Craton in southern Africa (Figure 2.4; Borg, 1988; Modie, 1996). The fold belt represents the southern foreland of the Damara Orogen, one of the many Pan-African mobile belts that
formed during the latest Neoproterozoic to Cambrian assembly of Gondwana (Martin and Porada, 1977a, b). The Damara fold belt overprints Mesoproterozoic volcanic and intrusive rocks and a thick package of unconformably overlying metasedimentary rocks interpreted to have been deposited in a rift-related environment; both igneous and sedimentary rocks within the belt are known to contain copper occurrences (Schwartz et al., 1995; Modie, 2000).

2.3.1 The Koras-Sinclair-Ghanzi rift system of southern Africa

The Kalahari Craton was spawned from a small composite Archaean core which grew by prolonged crustal accretion in the Paleoproterozoic along its NW side (Magondi–Okwa–Kheis Belt, Rehoboth Subprovince) to form the Proto-Kalahari Craton by ~1750 Ma (Figure 2.5; Jacobs et al., 2008). Between ~1400 to 1000 Ma, all margins of this crustal entity recorded intense tectonic activity: the NW margin was a major active continental margin between ~1400 to 1200 Ma and along the southern and eastern margins, the Namaqua–Natal–Maud–Mozambique Belt records a major arc–accretion and continent–collision event between ~1100 and ~1050 Ma. By ~1050 Ma, the Proto-Kalahari nucleus was
Figure 2.5 The incipient Kalahari Craton at ~1080 Ma. The Proto-Kalahari Craton was surrounded by Mesoproterozoic crustal additions, the active northwestern margin had stopped, the Umkondo-Borg large igneous province had been emplaced, and collisional and indentation tectonic processes (big arrow) were continuing along the southern margin of the Kalahari Craton. Whilst large amounts of juvenile crust had accreted along the SE, E, and NW margin, major components of older crust were accreted along the Namaqua Belt. Modified from Borg (1988) and Jacobs et al. (2008), see reference for further description.

almost completely rimmed by voluminous Mesoproterozoic crust and became a larger entity, the Kalahari Craton (Figure 2.5; Jacobs et al., 2008).

Immediately prior to the onset of arc–continent–continent collision along the Namaqua–Natal–Maud Belt (part of the widespread “Grenville-age” orogeny during which Rodinia was assembled), the Kalahari craton was subjected to intraplate magmatism – the Umkondo–Borg Large Igneous Province (LIP) – at ca. 1110 Ma (Figure 2.5; Jacobs et al., 2008). Several late middle Proterozoic volcano-sedimentary basins were developed along the western and northern margins of the Kalahari Craton, which were interpreted as two branches of a propagating continental rift system that extends for over 1000 km, the Koras-Sinclair-Ghanzi Rift (Figure 2.5; Borg, 1988). From south to north these are the Koras Basin
(South Africa), the Koras-Sinclair link, Sinclair, Klein Aub, and Dordabis/Witvlei basins in Namibia, and the Ghanzi/Lake N’Gami and Chobe (Goha and Chinamba Hills) basins in Botswana (Figure 2.5; Borg, 1988). The Ghanzi and Chobe basins were later grouped into the northwest Botswana rift based on the geophysical continuity of the basins (Key and Ayres, 2000). U-Pb zircon geochronology on the bimodal volcanic rocks within these basins indicate that the continental margin volcanism was contemporaneous with the ~1.1 Ga Umkondo-Borg LIP (e.g. Schwartz et al., 1996; Singletary et al., 2003; Hanson et al., 2004; Miller, 2012).

2.3.2 The Ghanzi basin

The Ghanzi basin straddles the border between eastern Namibia and western Botswana between the towns of Gobabis and Maun, respectively (Figure 2.6). It is comprised of folded bimodal metavolcanic rocks of the Kgwebe Formation overlain by meta-sedimentary rocks of the Ghanzi Group. The initial phase of rifting in the Ghanzi basin is represented by emplacement of bimodal volcanic rocks of the Kgwebe volcanic complex, which has a Pb-Pb zircon age of 1106 ± 2 Ma (Figure 2.7; Schwartz et al., 1996). The geochemistry of rhyolitic and basaltic rocks indicate that they formed through melting of Mesoproterozoic calc-alkaline rocks that were underplated in the middle and/or lower crust and were emplaced during late orogenic collision-associated extensional collapse (Kampunzu et al., 1998).

A subsequent phase of rift enlargement resulted in the deposition of the Ghanzi Group, a 5- to 7-km-thick sedimentary succession of continental to marine sedimentary rocks. From base to top, the Ghanzi Group is comprised of the Kuke, Ngwako Pan, D’Kar, and Mamuno formations (Figure 2.7). The basal Kuke Formation ranges from 0 to 500-meters-thick and consists of mature sandstone and immature conglomerate that contains clasts of Kgwebe Formation volcanic rocks. These rocks were interpreted to have been deposited in a fluvial environment with locally derived talus (Modie, 2000). The overlying Ngwako Pan Formation attains thickness of up to 3,500 meters. The lower portion of the formation consists of gray high mud-matrix sandstones while the upper portion of the formation consists of tan to pink (oxidized facies) sandstones with minor interbedded maroon mudstone and abundant parallel laminated ‘grit’ beds near the top of the formation (Modie, 1996, 2000). Although the Ngwako Pan Formation has been interpreted to have been deposited in a fluvial to shallow-marine environment, the regional context of the rocks is poorly documented due to scare outcrop and few boreholes; further work is needed to understand the regional depositional setting of these rocks (Lehmann et al., 2015).

Continued rift enlargement gave way to a marine incursion and deposition of the D’Kar Formation, which is the primary host to the copper deposits. It ranges from 1,500 to 2,500-meters-thick and consists of gray to green (reduced facies) mixed siliciclastic and carbonate rocks interpreted to have
Figure 2.6 Merged Total Magnetic Intensity map of the northern branch of the inverted Koras-Sinclair-Ghanzi rift system in Botswana and eastern Namibia, also known as the Kalahari Copperbelt. The belt is divided into two distinct aeromagnetic domains, the Rehoboth domain in Namibia and the Ghanzi-Chobe domain in Botswana (after Lehmann et al., 2015). The Ghanzi and Chobe basins are separated by the WNW-trending giant Karoo (Jurassic) Okavango dyke swarm.

Figure 2.7 Generalized lithostratigraphy of the Ghanzi-Chobe zone (Ghanzi basin) and interpretation of the sedimentary depositional environment. Modified from Modie (1996).
been formed in a shallow marine environment (Figure 2.7; Schwartz et al., 1995; Modie, 1996, 2000, Hall, 2013; Lehmann et al., 2015). Up to 1.5 km of purple-colored sandstones and interbedded carbonate rocks of the Mamuno Formation cap the Ghanzi Group sedimentary package (Figure 2.7). These rocks were interpreted to have been deposited in a near-shore to fluvial environment (Modie, 1996, 2000).

2.3.3 The Okwa Group and the Damara Orogen in Botswana

The deformed Ghanzi basin is separated from older Paleoproterozoic rocks of the Okwa Basement Complex by the Passarge Basin, which is filled with up to 10 kilometers of lacustrine to fluvialite sedimentary rocks of the Okwa Group (see Figure 2.4). These rocks were interpreted to have been deposited in a foreland basin setting to the Pan-African Damara Orogen (Ramokate et al., 2000). A youngest detrital zircon from the basal Takatswaane Formation has a U-Pb age of 579 ± 12 Ma, indicating that the Okwa Group was deposited in a foreland basin setting (see Figure 2.4) and received detritus from a variety of Paleo- to Neoproterozoic sources involved in the Pan-African Damara Orogen.

Continent-continent collision between the Kalahari and Congo cratons during the Damara Orogen resulted in inversion of the Proterozoic volcano-sedimentary basins and formation of a fold belt along the northern margin of the Kalahari Craton. The Ghanzi-Chobe zone (fold-and-thrust-belt) represents the southern foreland fold-and-thrust belt to the Damara Orogen (Modie, 1996). The limits of the fold belt have been defined through regional gravity and magnetics surveys (Figure 2.6; Hutchins and Reeves, 1980; Key and Ayres, 2000). More recently, the connection between the Ghanzi-Chobe zone and the Southern Foreland zone of the Damara Orogen were confirmed through the investigation of merged aeromagnetic datasets from Botswana and Namibia; the geology of the concealed fold belt was mapped for the first time based on the distinct magnetic fabrics within to the Kgwebe Formation and Ghanzi Group using three different filtered/processed grids (Total Magnetic Intensity, reduced-to-pole first vertical derivative, and reduced-to-pole total horizontal derivative grids; Figure 2.8; see following discussion on geophysical exploration for further details; Lehmann et al., 2015).

Progressive deformation in the region resulted in a 150-km-wide belt characterized by doubly-plunging anticline and syncline pairs with axial surfaces that can be traced over distances of 10 to 50 km and the axial traces anticline spaced 2 to 8 km apart across strike (Figures 2.8 and 2.9; Schwartz et al., 1995; Modie, 2000, Hall, 2013; Lehmann et al., 2015). Fold amplitudes are approximately 4-6 kilometers with the dip of fold limbs ranging between 45° to vertical; fold asymmetry defines northwest to southeast shortening for many of the folds, although exceptions to this have been noted (Figure 2.9; Schwartz et al., 1995; Veldsman, 2010; Hall, 2013). Satellite imagery over the Ghanzi Ridge area reveals that closely-spaced second-order parasitic folds are common within the D’Kar Formation (Hall, 2013). Fold axial planes strike 220° to 235° (right-hand-rule format) and dip between 80° to the northwest and
Figure 2.8 Merged Total Magnetic Intensity (TMI) aeromagnetic grids (top) and geological interpretation of the aeromagnetic data (middle) for the western portion of the Ghanzi-Chobe domain. Modified after Hall (2013) and Lehmann et al. (2015).
vertical. A penetrative northeast-trending axial plane cleavage is imparted to all formations within the Ghanzi-Chobe zone (Modie, 1996).

Deformation in the Southern Foreland zone in Namibia is broadly bracketed between 580 and 500 Ma based on Ar-Ar recrystallization ages from white micas, hornblende, and whole rock (Gray et al., 2006). Peak metamorphism and deformation occurred at roughly 530 Ma based on K-Ar ages from detrital white micas within the Nama Group in Namibia (equivalent to the Okwa Group; Horstmann et al., 1990). Post-peak deformation continued along major shear zones in the Kaoko and Damara Belts in Namibia through 460 Ma based on mica blocking temperatures and discordant Ar-Ar age spectra (Gray et al., 2006).

The deformed Ghanzi basin is cut by an array of strike-parallel to sub-parallel structures; one such structure was shown to have right-lateral displacement (Schwartz and Akanyang, 1994). A second set of north-northeast trending faults cut the folded Ghanzi Group rocks and typically display right-lateral displacement in map view; cross-cutting relationships indicate these faults offset the strike-parallel right-lateral-displacement fault system (Schwartz and Akanyang, 1996; Hall, 2013; Gill, 2016).

Structural analysis veins contained of the deformed Ghanzi Group rocks suggests that flexural-slip processes aided deformation within the belt (Figure 2.10). This is indicated by widespread bedding-parallel slip surfaces, bedding-parallel and discordant linking quartz-carbonate-(sulfide) vein arrays, and brittle-ductile shear fabrics within rheological weak, fine-grained lithologies (Figure 2.11). The rheological differences between interbedded fine- and coarse-grained lithologies facilitated slip along the limbs during fold amplification, resulting in brittle-ductile conditions within the fine-grained lithologies and brittle failure within the coarse-grained lithologies (Hall, 2013, Davies, 2013). Increased vein density (>5 veins per meter) are generally restricted to areas of contrasting rheology (i.e. the basal 10-15 meters
Figure 2.10 Schematic diagram of the flexural-slip mechanism and associated veins inferred to be representative of folding in the Ghanzi-Chobe zone. A) ‘Out-of-syncline’ flexural slip and tension gash arrays in a syncline. B) Combination of bedding-parallel veins and tension gashes. From Treagus (1992).

Figure 2.11 Examples of quartz-carbonate-(sulfide) veinlet arrays from the Boseto copper deposits. The dominant bedding-parallel orientation of the veinlets (images a-e) coupled with tension-gash and/or discordant veinlets (images a, d-f) with reverse shear sense indicators are indicative of the flexural-slip mechanism operating during deformation in the Ghanzi-Chobe zone. See Hall (2013) for full explanation.
of the Ngwako Pan-D’Kar contact. The syn-kinematic timing of the vein system is indicated by the presence of cleavage-parallel veins and reverse-shear sense structures including brittle offset of discordant veins along bedding planes, micro- to macroscopic folds, and brittle-ductile shear fabrics including boudinage of veins (Hall, 2013).

2.3.4 Post-Damara geology

The Ghanzi Chobe zone is cut by the giant Okavango dyke swarm; it forms a prominent feature in regional aeromagnetic surveys (Figure 2.6; Reeves and Hutchins, 1982). It consists of a group of east-southeast-trending Jurassic (~179 Ma) dolerite dykes (and some Proterozoic dykes that cross-cut older Paleoproterozoic crust) that cover a region that is approximately 2,000-km-long and 110-km-wide (LeGall et al., 2002; Jourdan et al., 2006). In addition to the dykes, the Ghanzi-Chobe zone is overprinted by discontinuous half-graben structures filled by sedimentary rocks (including glacial diamictites) and lavas of the Karoo Supergroup (Schwartz et al., 1995; Mackay, W., personal communication, 2012). The grabens are controlled by three fault sets: one that parallels the northeastern-trend of right-lateral faults that cut the Ghanzi-Chobe zone, a second that parallels the north-northeast trend of the late faults that cross-cut the belt, and a third east-southeast trending set that parallels the trend of the Okavango dyke swarm (Schwartz et al., 1995; Hall, 2013). The Karoo Graben #1 is situated ~15 km to the northwest of the Kgwebe Hills (northeastern Ghanzi Ridge), the margin of which contains Lake N’Gami.

The Ghanzi-Chobe zone and the Mesozoic Karoo Supergroup rocks are in turn overprinted by the nascent Okavango rift (Modisi et al., 2000). The faults bounding the incipient Okavango rift parallel those that cut the Ghanzi-Chobe zone and that bound the Karoo grabens, suggesting an inheritance of structural fabrics from the Ghanzi-Chobe zone (Modisi, 2000).

2.4 Geophysical exploration in the Kalahari Copperbelt

The Geological Survey of Botswana undertook nation-wide gravity surveys and high-resolution (250 m spatial resolution) magnetics surveys in the 1970’s and 1980’s to help interpret the geology underlying the vast aeolian sand cover of the Kalahari Desert (e.g. Figure 2.4; Hutchins and Reeves, 1980; Reeves and Hutchins, 1982; Key and Ayres, 2000; Singletary et al., 2003). Interpretations of the gravity surveys led to basic understanding of the regional basement geology of Botswana (e.g. Hutchins and Reeves, 1980; Yawsangratt, 2002; Chisenga, 2015). The magnetic surveys were conducted between 1976 and 1987 and were first interpreted in 1977, leading to the definition of the Ghanzi-Chobe zone in northwest Botswana (Hutchins and Reeves, 1980).

The regional geology of the Ghanzi-Chobe zone can be interpreted with high confidence when coupled with aeromagnetic data (Figure 2.8). This is due to the contrasting magnetic signatures of the Ghanzi Group rock formations. The Kgwebe Formation has an undulose texture with high intensity, low
frequency anomalies with ill-defined magnetic margins. The Ngwako Pan Formation of the Ghanzi Group has a flat texture with low intensity, low frequency anomalies with well-defined magnetic margins. The D’Kar Formation has an elongate texture with high intensity, moderate to high frequency anomalies with well-defined magnetic margins while the Mamuno Formation has a flat texture with medium to low intensity, low frequency anomalies (Lehmann et al., 2015). Aeromagnetic datasets from the governments of Botswana and Namibia were merged and interpreted on this basis, leading to the first detailed geologic map of the Kalahari Copperbelt and the Ghanzi basin (Figure 2.8; Lehmann et al., 2015). The datasets confirmed the existence of two distinct magnetic domains in the Kalahari Copperbelt, the Rehoboth domain and the Ghanzi-Chobe domain (Figure 2.8; Lehmann et al., 2015).

2.5 Exploration and development history of the Kalahari Copperbelt

Copper was first exploited in modern times from the Kalahari Copperbelt in 1966 at the Klein Aub mine, located south of Rehoboth in Namibia (Figure 2.12). The underground mine was closed in 1987 after producing 7.5 Mt @ 2% Cu and 45 g/t Ag (Borg and Maiden, 1987). Known copper occurrences in Klein Aub, Witvlei, and Dordabis areas of central and western Namibia (Figure 2.12) led Johannesburg Consolidated Investments, Anglovaal South West Africa, Pty., and U.S. Steel to extend regional copper exploration programs into western Botswana during the 1960s and 1970s. Exploration around the Lake N’Gami area resulted in the discovery of the Ngwako Pan copper deposit (now the Boseto copper deposits; Figure 2.12) with potential reserves of 20 million metric tons (Mt) of ore with a grade of 1.74% Cu and 39 pp Ag (U.S. Steel, 1978, in Schwartz et al., 1995). In the 1990’s, Anglo-American Corporation discovered the Banana Zone district, located ~60 kilometers to the southwest of the Lake N’Gami area. Further exploration in the belt was conducted from 1996 to 2000 by a joint venture involving Delta Gold of Zimbabwe (Delta), Kalahari Gold and Copper (Pty) Ltd of Namibia and Gencor/BHP Billiton. By 2010, the Banana Zone district had an inferred resource of 83 Mt averaging 1.42 percent Cu and 18.1 g/t Ag at a cut-off of 0.75% Cu. (Van Der Heever et al., 2010).

The Ngwako Pan copper deposit and surrounding properties were acquired by Discovery Metals Botswana, Ltd., in 2005. At the onset of production from the Zeta and Plutus mines in 2012 and 2013, respectively, the Ngwako Pan/Boseto copper deposits had a combined mineral resource (measured, indicated, and inferred) of 131.0 Mt at 1.3% Cu and 16.2 g/t Ag with a cut-off grade of 0.6% Cu and ore reserves of 29.1 Mt at 1.4% Cu and 19.8 g/t Ag (Discovery Metals Limited Report, 2012).

Cupric Canyon Capital and its African subsidiaries Cupric Africa and Khoemacau Copper Mining (Pty) Ltd. obtained the Ghanzi Project from Hana Mining Ltd. as well as the stalled Boseto open-pit operations (Discovery Metals Botswana) in 2013 and 2015, respectively. Exploration efforts have been focused on the Zone 5 deposit, a blind discovery under 40 meters of Kalahari Group cover that is located
approximately 35 kilometers south-southwest of the Boseto deposits. The Zone 5 stratiform Cu-Ag ore body (Figure 2.12) has a strike-length of 4,200 meters, dips at 60°, and averages 9-10 in true thickness; the deposit is open at depths greater than 1200 meters. The Zone 5 deposit has a current underground mineral resource of 100.8 Mt at 1.98% Cu and 20 g/t Ag using a 1% Cu cut-off grade (Cupric Canyon Capital report, 2017).

In addition to the Zeta and Plutus deposits and the Banana prospect, several shallowly-drilled prospects were discovered within a 60-km radius of the Boseto deposits by Discovery Metals and Hana Mining Ltd (Figure 2.12). Several satellite deposits have been discovered around the Boseto/Zone 5 districts including the Zeta Northeast (27 Mt @ 2.2% Cu and 40 g/t Ag with 1% Cu cut-off), Northeast Mango Two (21 Mt @ 1.7% Cu and 19 g/t Ag with 1% Cu cut-off), Zone 5 North (26 Mt @ 2% Cu and 40 g/t Ag with 1% Cu cut-off), and Zone 6 Cu-Ag deposits (17 Mt @ 0.9% Cu and 4 g/t Ag).

Exploration covering properties to the west and southwest of the Banana Zone resulted in the discovery of the Corner K/Mahumo copper-silver deposit (measured, indicated and inferred resources of 2.677 Mt @ 2% and 50 g/t Ag) and the T3 project (maiden resource of 23.36 Mt @ 1.24% Cu and 15.7 g/t Ag; Figure 2.12; MOD Resources, 2016).
2.6 Mineralization in the Kalahari Copperbelt

The base of the D’Kar Formation is known to be intermittently mineralized with copper sulfides over 220 km along the Ghanzi Ridge in Botswana (Schwartz et al., 1995). Disseminated to structurally-controlled cleavage lenticles and quartz-calcite-(sulfide) veinlets are hosted by gray and greenish-gray (reduced facies) siliciclastic rocks with composition of fine siltstone (grading to claystone) to coarse siltstone (grading to subarkose) and subordinate limestone (>50% calcite) and marlstone (33-50% calcite) that form a continuous layer at the base of the D’Kar Formation (Schwartz et al., 1995; Hall, 2013).

2.6.1 Deposit characteristics

Copper-silver deposits in the Ghanzi Ridge area of the Kalahari Copperbelt consist of stratabound disseminated to structurally-controlled ore-bodies that are 5- to 40-meters-thick and have strike lengths of 1.5 to 4 kilometers. The deposits display a chemical reduction-oxidation (redox) buffered mineral zoning consisting of hematite and barite within the oxidized footwall that passes up stratigraphy into an ore zone with chalcocite (used a group name), bornite, and chalcopyrite, and an outer halo of pyrite, sphalerite, and galena in reduced rocks (Figure 2.13; Schwartz et al., 1995; Hall 2013). Trace amounts of molybdenite, wittichenite (Cu₃BiS₅), and arsenopyrite have also been noted to occur along with the copper sulfide ore assemblage (Morgan, 2013; Shephard, 2014; Walsh, 2014; Piestrzynski et al., 2015). A less distinct lateral zonation of sulfides from northeast to southwest over a ~5 to 10 kilometers strike length was recognized at the Boseto copper deposits (Hall, 2013). Many of the deposits of the Kalahari Copperbelt are classified as Kupferschiefer-type sediment-hosted coper deposits.

More recently, copper sulfides have been discovered in up to three vertically stacked zones over 400 vertical meters of D’Kar Formation stratigraphy (T3 Project, MOD Resources Botswana/ Metal Tiger Plc. press release, 2017). Stacked mineralized zones were also discovered over the fold closure of a plunging anticline within the Ngwako Pan Formation at the Northeast Fold deposit of the Banana Zone (Figure 2.14). Vein- and fracture-hosted chalcocite, bornite, specular hematite, and molybdenite occur where they encounter altered (bleached) subarkose of the Ngwako Pan Formation (i.e. red-bed type) while veins crossing unaltered rock only contain specular hematite. Mineralized zones in the overlying D’Kar Formation displays typical stratigraphic redox patterns but chalcopyrite is the primary ore sulfide (unpublished reports, Khoemaça Copper Mining, 2013). Occurrences of structurally-controlled mineralized corridors within the D’Kar Formation have also been reported in the central and western portions of the Ghanzi basin (Ourea/T14 prospect, Botswana; Eiseb prospect, Namibia).

Disseminated sulfide minerals occur as mm-scale grains and aggregates of authigenic quartz, sulfides, calcite ± chlorite ± muscovite ± albite ± siderite with quartz- that are typically concentrated
Figure 2.13 Stratigraphic section from the Plutus copper-silver deposit displaying the vertical zonation in sulfide mineralogy that is typical of the sediment-hosted stratiform copper deposits of the Kalahari Copperbelt in Botswana. Modified from Hall (2013).
along lithological layering (see Figure 2.11c; Schwartz et al., 1995; Hall, 2013). Euhedral pyrite cubes up to 1-2 mm across and overgrow lithological layering are present throughout the stratigraphic section. Disseminated pyrite grains often display increasingly copper-rich (chalcopyrite, bornite, chalcocite) replacement rims (Borg and maiden, 1989, Schwartz et al., 1995; Hall, 2013).

Structurally-controlled copper sulfides occur in cleavage-lenticles, cleavage-parallel veinlets, discordant and bedding-parallel fractures, veinlets and veins, and in the matrix to brittle-ductile shear fabrics. Quartz-(calcite) cleavage lenticles and cleavage-parallel veinlets containing copper sulfide minerals suggest that copper mineralization was synchronous with deformation that imparted a vertical cleavage throughout the Ghanzi-Chobe zone (Schwartz et al. 1995; Modie, 2000). Most veins are oriented parallel to sub-parallel to bedding suggesting a strong lithological control on hydrothermal fluid flow. Subordinate discordant veinlets have orientations that structural analysis suggests form conjugate pairs that reflect the overall southeast-directed compressional stress regime during deformation (Hall, 2013). Bedding-parallel veinlets range in width from a few millimeters to ~10 cm on average, veins greater than one meter in width do occur, but are rather rare (see Figure 2.11; Schwartz et al., 1995).
The hydrothermal quartz-carbonate-sulfide vein arrays display cross-cutting relationships that indicate multiple veining and deformational events that affected the host rocks. Many discordant veins are displaced along lithological layering, implying bedding-parallel slip (see Figures 2.10 and 2.11). Copper sulfides that occur as films on the surfaces of bedding-parallel slip planes have slickenlines that indicate a reverse sense of displacement along bedding planes (Hall, 2013). In some instances, discordant veins that underwent boudinage are cross-cut by non-deformed veins of identical composition (Hall, 2013). Many bedding-parallel veins display crack-seal textures that suggest multiple pulse of fluid flow along the same horizons while others display cataclastic breccias with rotated clasts that indicate fluid flow was synchronous with slip along these planes (Figure 2.15; Davies, 2013; Hall, 2013). Copper sulfide often occur within pressure-shadows and shear fabrics associated with domains of ductile deformation. Automated mineralogy of brittle-ductile shear fabrics and quartz-carbonate-sulfide veining indicate that these structural features are often accompanied halos of minor silicification and potassic alteration (potassium feldspar and/or sericite) that, in some cases, overprints an albite-quartz-chlorite-sulfide assemblage (Figure 2.16).

2.6.2 Hydrothermal fluids

Fluid inclusion studies on mineralized quartz-calcite-sulfide veins from around the Ghanzi Ridge indicate that the hydrothermal fluids were likely metamorphic fluids. Reported homogenization temperatures for both primary and secondary fluid inclusions range from ~60° to ~300°C and salinities from 4 to 25 wt. percent NaCl-(CaCl$_2$) equivalent (Schwartz et al., 1995; Morgan, 2013; Shephard, 2014). Two fluid inclusion assemblages (FIAs) from different growth zones (inner and outer) in quartz crystals had respective homogenization temperatures of 165° to 190°C and 225° to 235°C and similar salinities of 15-20 wt.% NaCl-(CaCl$_2$) equivalent. The inner, equant to negative crystal shape fluid inclusion assemblage in quartz was interpreted to have been entrapped under lithostatic conditions while subsequent deformation-induced movement along the vein resulted in the opening of voids that caused a pressure drop, at which time the irregularly-shaped inclusions with carbonate solid inclusions were entrapped in the outer growth rim; calcite and chalcopyrite were precipitated within voids (J. Reynolds, personal communication, 2012, in Hall, 2013). Estimated entrapment temperatures for the fluid inclusions after pressure corrections were between 250° to 300°C, which is consistent with the observed lower greenschist facies mineral assemblage of the host rocks (Hall, 2013; Morgan, 2013; Shephard, 2014; Walsh, 2014). Comparison of fluid inclusion studies from various deposits is inferred to indicate a decrease in both homogenization temperature and salinity for non-cupriferous veins that occur distal to copper-rich zones (i.e. quartz-calcite-pyrite-sphalerite-galena; Walsh, 2014).
Crush-leach extraction data from vein-hosted quartz, calcite, and chalcopyrite display Cl/Br ratios that are similar to seawater, suggesting that the hydrothermal fluids inherited their Cl/Br signature from evaporated seawater (i.e. sedimentary brines; Hall, 2013). The data also indicated that the hydrothermal fluids were enriched in both Ca and Na with respect to most sedimentary brines such as those studied in the Central African Copperbelt. The fluid composition for the Boseto copper deposits is like that of fluids associated with metamorphic base metal deposits such as Coeur D’Alene (P. Emsbo, personal communication, 2012, reported in Hall, 2013).

2.6.3 Sulfur source

Sulfur stable isotopic analyses carried out on several of the deposits within the Kalahari Copperbelt display wide range of δ³⁴S values, ranging from +5 to -50‰ (Ruxton, 1986; Ruxton and Clemmey, 1986; Hall, 2013; Gorman, 2013; Shephard, 2014; Walsh, 2014). The majority of disseminated and structurally-controlled sulfide minerals (chalcocite, bornite, chalcopyrite), pyrite, and arsenopyrite have values that range from -10 to -30‰. Galena and sphalerite have δ³⁴S values that typically range from 0 to -25‰. Barite and euhedral pyrite are the only minerals reported with δ³⁴S values between +5 to 0‰ (Hall, 2013; Gorman, 2013; Shephard, 2014; Walsh, 2014). Systematic variations in δ³⁴S values were reported from the Boseto copper deposits; large negative shifts in δ³⁴S values occurred within the primary ores zones that are characterized by increased vein density as brittle-
Figure 2.16 Automated mineralogy analyses displaying overprinting veinlet generations from Boseto. Top: Two mineralogically distinct bedding-parallel veinlet sets (albite-quartz-bornite and calcite-dolomite-ankerite-quartz-bornite) and discordant veinlet sets (quartz-calcite-dolomite-ankerite-bornite and calcite-quartz). Bottom: A bedding-parallel shear plane cored by carbonate alteration overprints a bedding-parallel albite-quartz-chlorite-bornite and displaces a discordant quartz-calcite-bornite veinlet. An alteration selvage with silicification, potassic alteration (potassium feldspar and muscovite-biotite), and destruction of albite and chlorite is associated with the calcite-rich bedding-parallel veinlets in both top and bottom images. See Hall (2013) for complete description.
ductile deformation features (Figure 2.17; Hall, 2013). This was interpreted to reflect possible fractionation of sulfur isotopes during hydrothermal fluid flow (Hall, 2013). Overall, the sulfur stable isotopic data overwhelmingly point towards a reservoir of reduced sulfur that was probably derived through bacteriogenic reduction of sulfate (i.e. bacteriogenic pyrite within the host rocks).

2.6.4 Timing of mineralizing events

Although structural observations and data suggest that most of the copper in the Kalahari Copperbelt was introduced during deformation, the exact timing of the mineralizing was unknown prior to the work for this study. Whole-rock Pb and U isotopic data obtained from mineralized and non-mineralized rocks at the Klein Aub mine in Namibia pointed towards a ca. 600 Ma epigenetic mineralizing event. However, some indications were found of another, earlier and possibly syn- or diagenetic mineralizing event (Walraven and Borg, 1992).

Rhenium-Osmium (Re-Os) geochronology performed on vein-hosted chalcopyrite sample from the Plutus deposit returned an age poorly constrained range of 442 to 496 Ma (Hall, 2013), although the more accurate age could have been obtained by using the isochron method. Nonetheless, the age range points to a mineralizing event during the latest phases of Damara deformation (Hall, 2013). However, samples from the nearby and more deformed Zeta deposit returned Re-Os ages of 914 ± 4 Ma and 1012 ±

Figure 2.17 Sulfur stable isotope results by sulfide mineral and location (in meters) above the Ngwako Pan Formation (NPF) – D’Kar Formation contact. The range of $\delta^{34}$S values (0 to -45) is typical for deposits of the Kalahari Copperbelt. Modified from Hall (2013).
17 Ma for vein-hosted samples of low-level, highly radiogenic (LLHR) chalcopyrite and moderately LLHR bornite, respectively. Together with sulfur stable isotopic data, these Re-Os ages were interpreted to reflect incorporation of diagenetic pyrite into the vein system where the Re-Os isotopic signature of the pyrite was not re-set during recrystallization into copper sulfide minerals (Hall, 2013). The Zeta Re-Os sample are the oldest reported radiogenic ages for the Ghanzi Group.

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3.1 Abstract

New igneous and detrital zircon laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS) U-Pb geochronology and Lu-Hf isotopic data are presented for the Mesoproterozoic Kgwebe Formation and the unconformably overlying Meso- to Neoproterozoic Ghanzi Group in Botswana, the latter of which hosts sedimentary rock-hosted copper-silver deposits of the Kalahari Copperbelt. The age of deposition of the Ghanzi Group was previously constrained to between ~1100 and ~715 Ma, a period encompassing both the amalgamation and break-up of Rodinia, the latter of which has been shown to have important implications for the metallogeny of other Neoproterozoic sedimentary rock-hosted copper districts.

A Kgwebe Formation porphyritic rhyolite flow from the Makgabana Hills area of the Ghanzi Ridge yielded a weighted average $^{238}\text{U}/^{206}\text{Pb}$ zircon age of 1085.5 ± 4.5 Ma. This age provides a new maximum depositional age for the unconformably overlying Ghanzi Group.

Detrital zircon (n = 448) from stratigraphic positions below, within, and above the cupriferous D’Kar Formation have a $^{207}\text{Pb}/^{206}\text{Pb}$ age distribution with a dominant (70 to 90%) Mesoproterozoic (~1450 to ~1050 Ma) and a smaller (5 to 20%) Paleoproterozoic (~2200 to ~1700 Ma) population; a few grains (n = 4) are older (~3000 Ma to ~2450 Ma). Statistically, each sample contained a youngest cluster of overlapping zircon ages (n ≥ 3 at 2σ) that yielded weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages between ~1060 to ~1050 Ma, providing a maximum depositional age constraint on the D’Kar Formation host rocks.

Initial hafnium ratios ($\epsilon$Hf) for the Paleoproterozoic zircon populations vary between +7 and -12 with corresponding crustal residence model ages (from depleted mantle, $T_{\text{DM}}^C$) between ~3000 and 2050 Ma, indicating either fractionation from a chondritic uniform reservoir (CHUR) or mixing between juvenile mantle and older crustal components during magma generation. $\epsilon$Hf values for the Mesoproterozoic zircon populations vary between +15 to -18 and $T_{\text{DM}}^C$ model ages between ~3000 and...
~1100 Ma, suggesting that the source terrane(s) contained magmatic rocks including both juvenile rocks and substantially reworked Paleoproterozoic and possibly Archean crust.

Detrital zircon U-Pb and Lu-Hf data were compared to a compilation of published U-Pb, Lu-Hf, and Sm-Nd isotopic data for magmatic rocks from the Kalahari craton. The data suggest that the dominant Mesoproterozoic zircon population(s) were derived from the ~1450 to ~1000 Ma Rehoboth Subprovince and the Namaqua Sector of the Namaqua-Natal-Maud-Mozambique belt; more juvenile ~1200 to ~1000 Ma magmatic rocks of the Natal Sector and the Maud and Mozambique belts are unlikely sources for the detrital zircon. Sediment transport from the high-grade continent-continent collisional Namaquan-Grenvillian front was likely directed towards the northwest and north along the orogenic front before being re-directed towards the northeast, towards and the Ghanzi Basin, along with sediments from the Rehoboth Subprovince.

3.2 Introduction

The ~1110 Ma Kgwebe Formation bimodal volcanic sequence and the unconformably overlying Ghanzi Group in northern Botswana and easternmost Namibia comprise a volcano-sedimentary rift basin in the southwestern half of the northwest Botswana rift (Figure 3.1; Schwartz et al., 1996; Key and Mapeo, 1999; Singletary et al., 2003). The northwest Botswana rift forms the northeastern extension of the extensive Meso- to Neoproterozoic Koras-Sinclair-Ghanzi rift system along the northern and western margins of the Kalahari Craton (Figure 3.1; Borg, 1988; Hoffmann, 1989; Aldiss and Carney, 1992; Modie, 1996; Key and Ayres, 2000). The rift system was divided into several basins including the Koras basin in South Africa, the Sinclair, Klein Aub, and Dordabis/Witvlei basins in Namibia, and the Ghanzi/Lake N’Gami basin in Botswana (Figure 3.1; Borg, 1988), and the latter is the focus of this study (Figure 3.2). These basins host widespread copper-silver occurrences and some economic deposits, leading to the term Kalahari Copperbelt as a broad descriptor of the rift system along the northern margin of the craton (Borg and Maiden, 1989).

The Ghanzi Group is unconformably overlain by syn- to post-tectonic metasedimentary rocks of the Okwa Group in Botswana (Figure 3.2). The timing of deposition of these rocks is constrained by a U-Pb detrital zircon age of 579 ± 12 Ma (Ramokate et al., 2000). This broadly constrains deposition of the Ghanzi Group to between 1106 ± 2 Ma, by the U-Pb zircon age of the underlying Kgwebe Formation rhyolites (Schwartz et al., 1996; Figure 3.1), and the basin inversion during the Pan-African (~600 to ~480 Ma) Damara Orogen. Further age constraints come from correlations to with units in Namibia. The Ghanzi Group is correlated with the Tsumis Group in Namibia, which is in turn overlain by the Nosib, Witvlei, and the Nama groups (Damara Supergroup), where the latter is the equivalent of the Okwa Group.
Figure 3.1 Generalized Precambrian geology of the Kalahari Craton displaying the location of the Koras-Sinclair-Ghanzi rift system. The northeast trending branch of the rift is overprinted by Pan-African (600 to 480 Ma) deformation. The northwest Botswana rift is overprinted by the Ghanzi-Chobe zone, the southern foreland fold-and-thrust belt to the Damara Orogen. Modified after Hanson et al. (2004), Jacobs et al. (2008); and de Kock et al. (2014).

(Figure 3.3; Hegenberger, 1993; Miller, 2008; Lehman et al., 2015). A chemostratigraphy age correlation ties the Witvlei Group to Cryogenian ‘Snowball Earth’ glaciations based on the presence of thick diamictite layers and cap carbonate rocks with distinct trends in the $\delta^{13}$C isotopic composition that corresponding to the Sturtian (~717 to 660 Ma) and Marinoan (~635 Ma) glaciations (Hegenberger, 1993; Hoffman et al., 1998; Prave et al., 2011; Rooney et al., 2015). This suggests that the Ghanzi Group is older than ~717 Ma. No age data is available for the Nosib Group metasedimentary rocks.

This study presents new zircon U-Pb geochronology data for a rhyolite flow from the Makgabana Hills in the Ghanzi Ridge area as well as U-Pb and Lu-Hf geochronology and isotopic data for detrital zircon sampled from the western and central portions of the Ghanzi basin (Figure 3.2) to help refine the depositional age and provenance of the Ghanzi Group, particularly the cupriferous D’Kar Formation. The purpose of constraining the timing of deposition of the Ghanzi Group was to test whether it was deposited prior to or during the break-up of Rodinia, and in part to relate the metallogeny of Kalahari Copperbelt to
other significant Neoproterozoic basins containing sedimentary rock-hosted copper districts, such as the Central African Copperbelt in Zambia and Democratic Republic of the Congo (e.g. Hitzman et al., 2010). A better understanding of the provenance of the Ghanzi Group can provide insight into the tectono-stratigraphic evolution of the basin, and, when combined with geologic mapping, can help inform the search for suitable sedimentary, geochemical, and structural trap sites for hydrothermal copper-silver mineralization within the basin.

### 3.3 Geologic background

The Proto-Kalahari Craton consists of a core composed of two sutured Archean cratonic blocks, the Kaapvaal and Zimbabwe cratons, and adjacent Paleoproterozoic terranes (Jacobs et al., 2008). The composite Kalahari Craton includes Mesoproterozoic terranes in addition the Proto-Kalahari Craton (Figure 3.1; Jacobs et al., 2008). The extensional volcano-sedimentary basins of the Mesoproterozoic Koras-Sinclair-Ghanzi rift system occur along the northern and western margins of the Kalahari Craton
Figure 3.3 Stratigraphic correlations of Meso- to Neoproterozoic successions in the Kalahari Copperbelt of Botswana and Namibia (modified after Lehmann et al., 2015). Stars indicate approximate stratigraphic position of the current (red) and previous (white) detrital zircon samples from the Ghanzi and Tsumis groups. Yellow star indicates stratigraphic position of an igneous zircon from the Kgwebe Formation.
Bimodal magmatism throughout the rift system has been dated between 1112 and 1094 Ma and was coeval with mafic magmatism related to the 1110-1104 Ma Umkondo large igneous province (LIP) event that affected extensive areas of the Kalahari Craton (Figure 3.1; Ruxton, 1980, 1981; Hegenberger and Burger, 1985; Borg, 1998; Borg and Maiden, 1989; Schwartz et al., 1996; Singletary et al., 2003; Hanson et al., 1998; de Kock et al., 2014).

The Ghanzi/Lake N'Gami basin forms the central portion of the ~600-km-long northeast-trending northwest Botswana rift that stretches from northeastern Botswana to the western border with Namibia (Figure 3.1; Key and Ayres, 2000). The basin was inverted into a fold and thrust belt during the Pan-African (~600 to ~480 Ma) Damara orogeny. This belt is referred to as the Ghanzi-Chobe zone in Botswana (Figure 3.2) and the Southern Foreland zone of the Damara Orogen in Namibia (Miller, 2008; Lehmann et al., 2015). Lehmann et al. (2015) utilized regional geophysics to redefine the width and lateral extent of the belt in Namibia and Botswana as two continuous magnetic domains: the Rehoboth and Ghanzi-Chobe domains. The magnetics data confirmed the correlation of stratigraphy across international borders for the first time (Figure 3.3).

### 3.3.1 Stratigraphy and evolution of the Ghanzi basin

In Botswana, the bimodal volcanic suite that forms the base of the volcano-sedimentary basin is collectively known as the Kgwebe Formation (Figure 3.4). It consists of within-plate low Ti-P continental tholeiites and post-orogenic, within-plate high-K rhyolites that were inferred to have been derived from the melting of Mesoproterozoic calc-alkaline basement rocks (Kampunzu et al., 1998). However, Sm-Nd isotopic analyses carried out on Kgwebe Formation volcanic and intrusive rocks by Singletary et al. (2003) indicated that both Mesoproterozoic and Paleoproterozoic rocks were involved in their petrogenesis. The Mabeleapodi Hills porphyritic rhyolite, located in the Ghanzi Ridge area, has a U-Pb zircon age of 1106 ± 2 Ma (Figure 3.2; Schwartz et al., 1996). Several other porphyritic rhyolites (i.e. Oorlogsende porphyry member, Figure 3.2) and granitic to gabbroic sub-volcanic intrusions (i.e. Xade and Tshane Complexes; Figure 3.2) within and immediately outboard of the northwest Botswana rift have been dated using ID-TIMS zircon analysis at ~1110 to ~1094 Ma (Hegenberger and Burger, 1985; Singletary et al., 2003; Hanson et al., 2004b).

Bimodal volcanism in the northwest Botswana rift was followed by a second phase of extension that accommodated 5 to 10 km of siliciclastic and subordinate carbonate rocks of the Ghanzi Group (Modie, 1996, 2000). The Ghanzi Group is formally divided into the Ngwako Pan, D’Kar, and Mamuno formations, in ascending order of stratigraphy (Modie et al., 1998; Figure 3.4). The basal 500 meters of the Ghanzi Group, the informal Kuke Formation, consists of sandstones with thin conglomeratic layers (Figure 3.4). Clasts of Kgwebe Formation rhyolite within the conglomerate layers indicate an erosional unconformity between the Kgwebe Formation and Ghanzi Group.
Figure 3.4 Generalized stratigraphic column and unit descriptions for the northwest Botswana rift (modified after Modie, 1996, 2000). Stars denote approximate stratigraphic position of U-Pb igneous (yellow) and detrital (red) zircon samples. M.D.A. = maximum depositional age.

<table>
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<th>Gp.</th>
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<th>Sample/Age</th>
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| Okwa      | Mamuno       | 579 ± 12 Ma M.D.A. based on detrital zircon ID-TIMS U-Pb age (Ramadatsi et al., 2000) | -UPPER: sandstone, shale, dolomite, limestone, and conglomerate  
-LOWER: conglomerate, sandstone, siltstone, mudstone |
|           | Mamuno-1     |                                                 | reddish-purple cross-stratified to wave rippled arkosic sandstone interbedded with siltstone, mudstone, and limestone |
| Ghanzi    | D'Kar        | EISDD-008_189 HA-251-D_505                      | -UPPER: interstratified grey and reddish oxidized subarkose, sandstone, siltstone, and minor carbonate  
-LOWER: grey-green reduced planar laminated siltstone, subarkose, arkose, sandstone, and claystone with minor carbonate layers and black shale |
|           |              | Cu-Ag deposits disconformity                    |                                                                             |
|           | Ngwako Pan   | GBLD-004_206                                    | -UPPER: red oxidized planar laminated to plane-bedded sandstones, cm- to dm-scale ripple cross-laminated facies with red mudstone intraclasts and granule-rich layers  
-LOWER: high-matrix grey sandstone, normally graded laminations, and dark mudstone intraclasts |
|           |              |                                                 |                                                                             |
|           | K'uke        |                                                 | -grey quartz-arenite sandstone and red sandstone with thin conglomerate layers containing volcanic clasts |
|           |              |                                                 |                                                                             |
|           | Kgwebe       | DMDD2183_101                                    | -porphyritic metarhyolite flows with interbedded metabasalt and metasedimentary rocks |
|           |              | 1106 ± 2 Ma ID-TIMS U-Pb age (Schwartz et al., 1996) |                                                                 |
|           |              |                                                 |                                                                             |
|           | Okwa Btmt. Clmplx. | 2056 ± 2 Ma ID-TIMS U-Pb age (Modie et al., 2006) | -felsic volcanic and plutonic rocks |


The Ngwako Pan Formation is approximately 2000- to 3500-meters-thick in the Ghanzi Ridge area (Schwartz et al., 1995; Modie, 2000). The basal part consists of gray colored wackestone overlain by well-sorted red sandstones and arkoses that are locally interbedded with pebbly layers and granulestone (Modie, 1996). The upper part of the formation is characterized by the predominance of planar laminated plane-bedded sandstones together with cm- to dm-scale ripple cross-laminated facies containing rip-up clasts of mudstone and associated graded beds of granulestone (Modie, 1996). Recent investigations of high-resolution aeromagnetic data in the Ghanzi Ridge area indicate that the Ngwako Pan Formation is characterized by rapid thickness changes, pinching to a few hundred meters or less in places above presumed paleotopographic highs formed by the Kgwebe Formation (Hall and Hitzman, 2016). Modie (1996) suggested that the Kuke and Ngwako Pan formations were deposited in a continental fluvial depositional system that followed the axis of the northwest Botswana rift. However, Lehmann et al. (2015) noted that the Ngwako Pan Formation rocks lack sedimentary structures typically associated with braided to meandering fluvial systems and suggested that the paleo-environmental setting of the Ngwako Pan Formation requires further investigation.

The second stage in the development of the Ghanzi basin involved marine transgression and deposition of the D’Kar Formation. The contact between the D’Kar and Ngwako Pan formations has previously been described as conformable (Huch et al., 1992; Schwartz et al., 1996; Modie, 1996). However, mapping of continuous magnetic lithostratigraphic anomalies within Ngwako Pan Formation revealed an angular unconformity between the Ngwako Pan and D’Kar formations in the vicinity of the Ghanzi Ridge (Hall and Hitzman, 2016). The data suggest that uplift and erosion of the upper members of the Ngwako Pan Formation occurred prior to marine transgression; this may have resulted in cannibalization and re-sedimentation of the sedimentary rocks of the uppermost Ngwako Pan Formation (Hall and Hitzman, 2016).

The D’Kar Formation is divided into lower and upper members Figure 3.4; (Schwartz et al. 1995). The lower member is characterized by reduced facies (gray to greenish-gray), parallel laminated siliciclastic metasedimentary rocks (siltstone, subarkose, arkose, sandstone, and claystone) and subordinate limestone/marlstone with or without dolomite (Schwartz et al., 1995). Dark gray to black colored siltstone beds have been interpreted as black mudstone or shale (Lehmann et al., 2015); however, total organic carbon contents are generally below 1%. Schwartz et al. (1995) suggested that the organic carbon content of these rocks was reduced during post-depositional hydrothermal fluid flow events. The upper member consists of interstratified, gray reduced facies and reddish oxidized facies metasedimentary rocks, which have the composition of subarkose, sandstone, and siltstone (Schwartz et al., 1995). Modie (1996) interpreted this stage of basin development to have resulted from thermal subsidence following the deposition of the Ngwako Pan Formation.
The uppermost portion of the fill constitutes the Mamuno Formation, which is approximately 1.5- to 2-km-thick (Figure 3.4; Schwartz et al., 1995; Modie, 1996). The contact between the D’Kar and Mamuno formations was described as a disconformity (Schwartz and Akanyang, 1994). The Mamuno Formation is composed of dominantly red-purple (chemically oxidized) well-sorted, fine- to medium-grained arkosic sandstone interbedded with siltstone, mudstone, and limestone characterized by planar lamination, cross-stratification, reactivation surfaces overlain by massive beds, ladder-back interference oscillatory ripples and straight-crested symmetrical ripple forms (Litherland, 1982; Modie, 1996). These rocks represent a transition from a shallow shelf environment to a nearshore/supratidal and/or subaerial fluvial-deltaic environment as accommodation space in the marine basin was filled (Schwartz et al., 1996; Modie, 1996).

The entire Ghanzi Group was initially correlated with the Nosib Group in the Damara Belt in Namibia (Litherland, 1982; Germs, 1995). More recently, only the Mamuno Formation has been tentatively correlated with the lower Nosib Group, the Kamtsas Formation (Figure 3.3; Schalk, 1970; Killick, 1983; Hoffmann, 1989; Schwartz et al., 1996; Ramokate et al., 2000; Miller, 2008).

3.3.2 Previous detrital zircon studies

Kampunzu et al. (2000) obtained ion microprobe U-Pb ages for detrital zircons from rock interpreted to be from the lowermost D’Kar Formation in borehole CPK4, located near the Goha Hills (Chobe basin) in northern Botswana (Figure 3.3). However, Meixner and Peart (1984) interpreted these metasedimentary rocks as a part of the Kgwebe Formation based on intercalated rhyolite layers. Detrital zircons yielded a main population with a range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1124 ± 15 and 1080 ± 21 Ma and a weighted-mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1104 ± 16 Ma (MSWD = 1.6, n = 12; Kampunzu et al., 2000). Statistically, the ages of the detrital zircons from this population were similar suggesting a single source provenance. The grains were interpreted to have been locally sourced from the ~1106 Ma Kgwebe Formation. Three other detrital grains yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1363 ± 11, 1759 ± 13, and 1736 ± 12 Ma. This early Mesoproterozoic zircon was interpreted to have been sourced from the Choma-Kalomo Block in Zambia, or possibly the Rehoboth basement/Namaqua Belt in Namibia. The source of the older Paleoproterozoic zircon was suggested to be Paleoproterozoic terranes in northern Namibia and southern Angola.

Zircons from three core samples of a supposed volcanic tuff/breccia horizon within the lower D’Kar Formation, obtained from the Zeta Cu-Ag mine (Figures 3.2 and 3.3) in the central Ghanzi basin, were analyzed by Talavera (2012). The combined analyses (n = 27) yielded a large $^{206}\text{Pb}/^{238}\text{U}$ age population (n = 16) with an age range from 1393 ± 17 to 1042 ± 13 Ma, a minor age population (n = 10) with a range from 1951 ± 26 to 1721 ± 27 Ma, and one age of 2246 ± 25 Ma (Talavera, 2012). The results are inconsistent with a volcanic source for the rock and the horizon has been reinterpreted by mine
geologists as a strongly deformed siliciclastic horizon (W. MacKay, personal communication, 2013).
The reported ages are consistent with provenance from a dominantly middle to late Mesoproterozoic
source terrane with lesser input from Paleoproterozoic sources.

In the western Ghanzi basin, Gill (2016) analyzed detrital zircons from the footwall (Ngwako Pan Formation) and the hanging-wall (D’Kar Formation) to a copper-silver occurrence (Figure 3.3). The footwall sample yielded a dataset (n = 121) that contained a distinct distribution with Mesoproterozoic $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from ~1400 to ~1000 Ma and Paleoproterozoic $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from 2000 to 1600 Ma, and two Neoarchean grains (Gill, 2016). The hanging-wall succession yielded a similar distribution with Mesoproterozoic $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging between ~1500 and ~1000 Ma and Paleoproterozoic ages between ~2200 and ~1600 Ma (Gill, 2016).

Steven et al. (2015) analyzed detrital zircons obtained from green siltstones at the top of the Eskadron Formation (Ngwako Pan equivalent, Figure 3.3) within the Witvlei sub-basin in east-central Namibia (Figure 3.2). The sample yielded two $^{207}\text{Pb}/^{206}\text{Pb}$ age peaks at 1248.9 ± 9.3 Ma (n = 12; MSWD = 0.40) and 1874.2 ± 7.2 Ma (n=12; MSWD = 0.42) and two Archean grains at ~3200 and ~2800 Ma. The youngest concordant zircon had a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1015 ± 26 Ma (Stevens et al., 2015).

### 3.4 Samples

One igneous zircon sample and four detrital zircon samples were collected for geochronological analyses. Drill hole DMDD2183 from the Makgabana Hills (Table 3.1, Figure 3.2 to 3.4) intersected intercalated massive porphyritic volcanic rock and volcaniclastic rocks of the Kgwebe Formation that consisted of rhyolite with an aphanitic groundmass and abundant magmatic phenocrysts, small to large, irregularly shaped, porphyritic blocks and clasts, and injections of metasedimentary rocks. A least altered portion of the upper massive porphyritic rhyolite was sampled for U-Pb dating.

U-Pb ages and Hf isotopic data were obtained from detrital zircon from three exploration drill holes and an outcrop located within the Ghanzi basin (Table 3.1; Figure 3.2-3.4). The four samples selected for detrital zircon analysis were representative of the uppermost Ngwako Pan Formation, the basal and lower (upper?) D’Kar Formation, and the basal portion of the Mamuno Formation. Two samples were obtained from the western Ghanzi basin in eastern Namibia while the other two samples were collected from the central Ghanzi basin in Botswana (Table 3.1).

### 3.5 Methods

Approximately 2 kg of Makgabana Hills rhyolite was collected from drill core and prepared for analysis using standard crushing and separation techniques, including heavy liquid and magnetic separation at the USGS Isotope Research Laboratory in Denver, CO. Zircon grains were handpicked under a binocular microscope and mounted in epoxy and polished. Cathodoluminescence (CL) imaging of individual zircon grains was used to characterize zoning and presence of inclusions (see Supplemental
Table 3.1 Sample locations for igneous and detrital zircon U-Pb and Hf isotope analyses

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Type</th>
<th>Formation</th>
<th>Location</th>
<th>UTM zone (WGS84)</th>
<th>Northing (m)</th>
<th>Easting (m)</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>DMDD2183_101</td>
<td>igneous</td>
<td>Kgwebe</td>
<td>Makgabana Hills</td>
<td>34S</td>
<td>7707077</td>
<td>712563</td>
<td>Gray, rhyolite with white/pink feldspar phenocrysts, irregular flow foliation</td>
</tr>
<tr>
<td>GBLD-004_206</td>
<td>detrital</td>
<td>Ngwako Pan</td>
<td>Farm #839, Namibia</td>
<td>34S</td>
<td>7612695</td>
<td>420227</td>
<td>Pink, massive arkosic sandstone, 35 m below Ngwako Pan-D’Kar fm. contact</td>
</tr>
<tr>
<td>EISDD-008_189</td>
<td>detrital</td>
<td>D’Kar</td>
<td>Fortuna Farm, Namibia</td>
<td>34S</td>
<td>7591196</td>
<td>382062</td>
<td>Brown-gray, cross-stratified sandstone with heavy mineral laminations interbedded with argillite, 55 m above Ngwako Pan Fm. contact</td>
</tr>
<tr>
<td>HA-251-D_505.4</td>
<td>detrital</td>
<td>D’Kar</td>
<td>New Discovery prospect, Botswana</td>
<td>34S</td>
<td>7666515</td>
<td>673241</td>
<td>Gray, thin-bedded sandstone with heavy mineral laminations, 180 m above Ngwako Pan Fm. contact</td>
</tr>
<tr>
<td>Mamuno-1</td>
<td>detrital</td>
<td>Mamuno</td>
<td>Zone 9 prospect, Botswana</td>
<td>34S</td>
<td>7707345</td>
<td>752483</td>
<td>Outcrop sample, purple-brown arkosic sandstone with minor intercalated limestone layers, lower Mamuno Fm.</td>
</tr>
</tbody>
</table>

Electronic File), and was performed on a JEOL 5800 scanning electron microscope at the USGS Microbeam Laboratory, Denver, CO.

Analyses of igneous zircon grains were conducted using a Nu Instruments Atto™ LA-SC-ICPMS at the USGS Southwest Isotope Research Laboratory in Denver, CO. Zircon was ablated with a Photon Machines Excite™ 193 nm ArF excimer laser in spot mode (150 total bursts for zircon) with a repetition rate of 5 Hz, laser energy of ~3 mJ, and an energy density of 4.11 J/cm². Pit depths are typically less than 20 µm. The rate of He carrier gas flow from the HelEx cell of the laser was ~0.6 L/min. Make-up Ar gas (~0.2 L/min) was added to the sample stream prior to its introduction into the plasma. Nitrogen with flow rate of 5.5 mL/min was added to the sample stream to allow for significant reduction in ThO+/Th+ (<0.5%) and improved the ionization of refractory Th (Hu et al.,2008). The laser spot sizes for zircon were ~25 µm. With the magnet parked at a constant mass, the flat tops of the isotope peaks of \(^{202}\text{Hg}, \ ^{204}\text{(Hg+Pb)}, \ ^{206}\text{Pb}, \ ^{207}\text{Pb}, \ ^{208}\text{Pb}, \ ^{232}\text{Th}, \ ^{235}\text{U}, \ ^{238}\text{U}\) were measured by rapidly deflecting the ion beam with a 30 s on-peak background measured prior to each 30s analysis. Raw data were reduced off-line using the lolite™ 2.5 program (Paton et al., 2011) to subtract on-peak background signals, correct for U-Pb downhole fractionation, and normalize the instrumental mass bias using external mineral reference materials, the ages of which had previously been determined by ID-TIMS. Ages were corrected by standard sample bracketing with the primary zircon reference material Temora2 (417 Ma; Black et al., 2004) and secondary reference material Plešovice (337 Ma, Sláma et al., 2008) and an in-house standard WRP-63-08 (1707 Ma; W. Premo, pers. comm., 2016). Reduced data were compiled into Wetherill concordia diagrams using Isoplot 4.15 (Ludwig, 2012). \(^{206}\text{Pb}/^{238}\text{U}\) ages are reported for igneous zircon samples less than ~1300 Ma and \(^{207}\text{Pb}/^{206}\text{Pb}\) ages are used for older ages following the recommendations of Gehrels (2012).
Sedimentary rock samples utilized for detrital zircon U-Pb and Lu-Hf analyses were crushed, pulverized, and sieved to obtain the 32- to 200-mesh fraction. To concentrate zircon crystals, heavy liquid separation was done in methylene iodide (ρ = 3.32 g/cm³). The “heavies” were briefly washed in acetone, followed by a final Frantz magnetic separation at ~1.5 mA. Approximately 140-160 randomly selected grains were handpicked from each sample under a binocular microscope for inclusion onto an epoxy mount. The mount was ground to expose the approximate center of the grains, polished, and then imaged with reflected light and CL to identify internal structures, zoning related to chemical composition, and defects such as fractures and metamict zones (see Supplemental Electronic File). CL imaging was performed on a JEOL 5800 scanning electron microscope at the USGS Microbeam Laboratory, Denver.

To obtain the lowest analytical uncertainty the same spots were analyzed separately for U-Pb and Lu-Hf isotopes with spot sizes of 20 μm and 50 μm, respectively. U-Pb and Lu-Hf isotopic analyses of detrital zircon were conducted on a Nu Instruments Plasma HR LA-MC-ICPMS instrument at the University of California Santa Barbara. Analyses were corrected for well-known isobaric interferences and mass bias corrections following the analytical procedures of Kylander-Clark et al. (2013). The U-Pb data were corrected against the 91500 zircon reference material (Wiedenbeck et al., 1995), and GJ-1, Plešovice, 91500, and R33 reference materials (Black et al., 2004; Woodhead and Hergt 2005; Blichert-Toft, 2008; Morel et al., 2008; Sláma et al., 2008) were used to ensure accuracy; these secondary reference materials yielded dates within 2% of their accepted values. Some data from metamict grains and/or grains that contained high common lead contents were rejected during data reduction and acquisition.

Following the method proposed by Gehrels (2012), all U-Pb data that was >10% discordant (discordance = 1 – 100*([206Pb/238U age]/[207Pb/206Pb age])), >5% reversely discordant, or had an error >10% of the measured 207Pb/206Pb age was rejected from further analysis and statistical comparisons. Accepted and rejected U-Pb data are included in the Supplemental Electronic File. Tera-Wasserburg concordia plots, relative age probability plots with stacked histograms, and weighted-mean age calculations were carried out using Isoplot 4.1 (see Supplemental Electronic File; Ludwig, 2012). For detrital zircon, 207Pb/206Pb ages are reported for grains greater than 1000 Ma following the methodology outlined by Gehrels (2000) and Gehrels et al., (2008) that suggests using a cut-off near ~1200 Ma that does not artificially divide a cluster of analyses as well as for consistency with other detrital zircon studies performed in the region.

Lu-Hf isotopic analysis was performed on selected grains with concordant U-Pb data and growth zones (determined from CL images) that were large enough to contain the additional Lu-Hf laser pit. Lu-Hf analyses were also conducted on the Nu Instruments Plasma LA-MC-ICPMS. Reference materials GJ-1, Mud Tank, Plešovice and 91500 zircon standards (Wiedenbeck et al., 1995, 2004; Woodhead and
Hergt 2005; Blichert-Toft, 2008; Morel et al., 2008; Sláma et al., 2008) were run to ensure accuracy, and yielded results equivalent to accepted values. Lu-Hf isotopic data tables are included in the Supplemental Electronic File. The $^{176}\text{Lu}$ decay constant $\lambda = 1.867 \times 10^{-11} \text{y}^{-1}$ (Söderlund et al., 2004) was used for all Hf isotope calculations. Present-day chondritic values of $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ (Bouvier et al., 2008) were used for calculation of initial hafnium ($\varepsilon\text{Hf}_i$) values and the chondrite uniform reservoir (CHUR) curve. Calculated epsilon hafnium values at the time of crystallization ($\varepsilon\text{Hf}_t$) are plotted against the corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ age date for each spot analyzed. This allowed the data to be compared to curves projecting the isotopic composition of mantle and chondritic reservoirs through Earth history as well as the potential evolution of the source material from which the detrital zircon grains crystallized (e.g., mantle or crustal reservoirs).

A modified (Andersen et al., 2011) version of the depleted mantle model of Griffin et al. (2000) was used. This produces a present-day $\varepsilon\text{Hf}_i = +16.4$ similar to that of average MORB over 4.56 Ga starting from chondritic initial hafnium. There are several approaches to calculating Hf model ages. Depleted mantle model ages ($T_{DM}$) ages are calculated using the measured (i.e., present-day) $^{176}\text{Lu}/^{177}\text{Hf}$, and $^{176}\text{Hf}/^{177}\text{Hf}$, of the zircon sample and model depleted mantle with a present-day $^{176}\text{Hf}/^{177}\text{Hf}_{DM} = 0.28325$ and $^{176}\text{Lu}/^{177}\text{Hf}_{DM} = 0.0388$ for projecting the Hf-isotope evolution back onto the depleted mantle curve. This gives a minimum age for the source material of the magma from which the zircon crystallized (Griffin et al., 2002). Therefore, we used a two-stage model to calculate the crustal residence (from depleted mantle) model age ($T_{DMC}$). This required a back-calculation to the present-day Hf isotope ratio ($^{176}\text{Hf}/^{177}\text{Hf}_0$) from the calculated initial Hf isotope ratio ($^{176}\text{Hf}/^{177}\text{Hf}_i$) utilizing a slope equal to $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$. This method assumes that the magma from which the zircon grew was produced by melting of average continental crust. The crustal residence age is then calculated by projecting the present-day Hf isotopic ratio ($^{176}\text{Hf}/^{177}\text{Hf}_0$) onto the depleted mantle curve. The crustal residence age represents the time of zircon separation from the depleted mantle (Griffin et al., 2002). Lu-Hf depleted mantle model ages are highly dependent on the parameters used and should therefore not be interpreted as real ages, but rather should be used as broad indicators of the age of crust forming processes. Lu-Hf isotope data and calculated initial hafnium values, depleted mantle model ages ($T_{DM}$), and crustal residence model ages ($T_{DMC}$) are included in the Supplemental Electronic File.

3.6 Geochronology results

LA-ICPMS U-Pb zircon geochronology was successfully carried out on one igneous rock sample and four detrital zircon samples. Lu-Hf isotopic analysis was successfully applied to detrital zircon that could accommodate the spot size of the laser beam.
3.6.1 Porphyritic rhyolite, Makgabana Hills, east-central Ghanzi basin

Weakly altered massive rhyolite from the Makgabana Hills (DMDD2183_101) was sampled for U-Pb age dating. Many zircon grains from the sample were blocky to elongate, euhedral, doubly terminated crystals consisting of uniform cores surrounded by oscillatory growth zones indicating primary growth in a magmatic environment (Figure 3.5). No metamorphic rims were encountered. Thirty-three concordant spot analyses yield a $^{206}\text{Pb}/^{238}\text{U}$ age of $1085.5 \pm 4.5$ Ma (MSWD = 0.081; Table 3.2; Figure 3.5), which is taken as the age of igneous crystallization.

3.6.2 Detrital zircon U-Pb results

The sample of Ngwako Pan Formation coarse-grained sub-arkosic sandstone from the western Ghanzi basin (GBLD-004) yielded 133 zircon grains for analyses. Among these was a very small population of colorless to reddish-brown zircon that ranged from rounded to prismatic, with the largest grain measuring 350 by 100 µm. CL analysis revealed a variety of textures including regular oscillatory zoning with common light and dark banding, grains that exhibited fractures, very bright but poorly zoned grains, and grains that display evidence of recrystallization. LA-ICPMS analyses yielded 100 concordant (<10% discordance) U-Pb spot dates. The analyses define a large (71%) $\sim 1350$-$1000$ Ma $^{207}\text{Pb}/^{206}\text{Pb}$ age population (Figure 3.6). Approximately 20% of the concordant analyses define a minor age population between $\sim 1920$ and $\sim 1700$ Ma, while 5% of the concordant analyses define a small age population ranging in age from $\sim 2150$ to $\sim 2040$ Ma. The sample also yielded single grains with ages of $\sim 1502$ Ma, $\sim 1513$ Ma, and $\sim 2552$ Ma (not shown in figures to highlight the younger zircon populations).

Sample EISDD-008_189.4 from the D’Kar Formation in the western Ghanzi basin yielded a voluminous and heterogeneous population of pale pink to amber to pale brown zircon of multiple shapes and sizes. The zircon grains were noticeably larger than those in other samples from this study, with some exceeding 400 µm in length. CL analysis reveals that the majority of zircon have concentric oscillatory zoning, while a minority display less obvious zoning and appear uniformly bright. LA-MC-ICPMS analyses yielded 139 concordant U-Pb spot dates. The analyses define a dominant (80%) age population between $\sim 1425$ and $\sim 1030$ Ma accompanied by a minor (14%) age population between $\sim 1950$ and 1750 Ma and a smaller (5%) age population between $\sim 2150$ and $\sim 2000$ Ma (Figure 3.6). The analyses also yielded single grains with ages of 2576 Ma and 2462 Ma.

In the central Ghanzi basin, D’Kar Formation sample HA-252-D_505.4 produced a high yield of pale-pink to pale-violet colored, euhedral to sub-rounded zircon, commonly 200-250 µm long. CL imaging revealed a range of textures from oscillatory to convoluted zoning, grains with minimal or no zoning dominated by dark patches, and grains with recrystallized margins. Analysis yielded 127 concordant U-Pb spot dates. The LA-ICPMS analyses define a dominant (87%) age population between $\sim 1380$ Ma and $\sim 1060$ Ma (Figure 3.6). Approximately 10% of the concordant analyses define a minor
<table>
<thead>
<tr>
<th>Sample</th>
<th>Concentrations</th>
<th>Isochrons</th>
<th>Error</th>
<th>Ages (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DMDD-01</td>
<td>99.9</td>
<td>2.049</td>
<td>0.049</td>
<td>1129</td>
</tr>
<tr>
<td>DMDD-02</td>
<td>167.2</td>
<td>0.58</td>
<td>0.001</td>
<td>787</td>
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<td>DMDD-03</td>
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<td>8.53</td>
<td>0.004</td>
<td>1061</td>
</tr>
<tr>
<td>DMDD-04</td>
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<td>0.005</td>
<td>1065</td>
</tr>
<tr>
<td>DMDD-05</td>
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<td>6.04</td>
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<td>DMDD-06</td>
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<td>0.003</td>
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<tr>
<td>DMDD-07</td>
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<td>5.47</td>
<td>0.004</td>
<td>1061</td>
</tr>
<tr>
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<td>0.003</td>
<td>1061</td>
</tr>
<tr>
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<td>93</td>
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<td>DMDD-33</td>
<td>76.6</td>
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Figure 3.5 U-Pb Wetherill concordia diagram for igneous zircon, Makgabana Hills porphyritic rhyolite. Inset: CL imagery displaying typical igneous zircon and locations of individual LA-ICPMS spot analyses.

age population at ~1960 Ma and ~1860 Ma and a smaller (2%) age population exists between ~2150 and ~2000 Ma. One concordant analysis yielded an age of ~1608 Ma.

The zircon yield from a sample of the Mamuno Formation outcrop (Mamuno-1) was limited. Recovered zircon are light pink and were often broken. Euhedral grains are small, rarely exceeding 200 µm in length. Examination under CL indicates that most grains exhibit well-defined and minimally disturbed oscillatory zoning. Analysis yielded 86 concordant U-Pb spot dates. The LA-ICPMS analyses define a dominant (88%) age population ranging from ~1370 to ~1040 Ma, and a small age population between ~1910 and ~1760 Ma (7%) and ~2150 and ~2000 Ma (3%; Figure 3.6). One grain yielded a concordant Archean age of ~2926 Ma.

A relative probability plot of the combined Ghanzi Group dataset (n = 452) reveals a dominant Mesoproterozoic age population between ~1425 and ~1000 Ma, a minor Paleoproterozoic age population between ~2150 to ~1700 Ma, and minor Archean and ~1600 to 1500 Ma age populations (Figure 3.7). The Archean grains of ~2925, ~2570, and ~2460 Ma are not displayed in the cumulative probability plots in order to highlight the dominant late Paleoproterozoic and Mesoproterozoic populations. In individual plots (Figure 3.6), a Paleoproterozoic age cluster displays minor peaks between ~2150 to 2000 Ma (group 1). A younger ~1950 to ~1700 Ma age cluster (group 2) contains a prominent age population between ~1950 and ~1860 Ma (group 2a). A younger ~1820 to ~1700 Ma population (group 2b) is prevalent in
Figure 3.6 A-D) Relative probability plots with stacked histograms for each sample displaying >90% concordant data. E-H) Plots of initial hafnium (\(\varepsilon\text{Hf}_{(i)}\)) versus \(^{207}\text{Pb}/^{206}\text{Pb}\) age for individual samples. \(T^c_{\text{DM}}\) crustal residence model age projections back on to the depleted mantle curve are based on recycling of average continental crust with \(^{176}\text{Lu}/^{177}\text{Hf} = 0.015\). DM = depleted mantle reservoir curve, CHUR = chondritic uniform reservoir curve. The Archean grains are not displayed in the cumulative probability plots in order to highlight the dominant late Paleoproterozoic and Mesoproterozoic populations.
Figure 3.7 Top: Relative probability plot with histograms displaying all Ghanzi Group detrital zircon data. Bottom: Plot of initial hafnium ($e^{\text{Hf}}_i$) versus $^{207}\text{Pb}/^{206}\text{Pb}$ age. DM = depleted mantle reservoir curve, CHUR = chondritic uniform reservoir curve, $T^c_{\text{DM}} = $ crustal residence (from depleted mantle) model age projection. $T^c_{\text{DM}}$ projections constructed utilizing a slope equivalent to $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ assuming melting of average continental crust. The Archean grains are not shown in the plot to highlight the Paleoproterozoic and Mesoproterozoic populations.
the western Ghanzi basin samples and nearly absent from the eastern Ghanzi basin samples (Figure 3.6). Three ~1600 to 1500 Ma zircon grains comprise group 3 (Figure 3.6). Individual relative probability plots of the ~1450 to ~1000 Ma age populations (group 4) display subsidiary age populations at ~1300 (group 4a), ~1200 Ma (group 4b), ~1110 Ma (group 4c), and ~1060 – 1050 Ma (group 4d; Figure 3.6).

Calculating a youngest detrital zircon age population from a set of detrital zircon analyses can be difficult due to potential analytical uncertainty, Pb loss, inheritance, and the presence of grains with a range of ages (Dickinson and Gehrels, 2009; Gehrels, 2014). While the Mesoproterozoic detrital zircon age population from this study contains overlapping U-Pb ages with uncertainties (at 95% confidence) that form a continuous distribution, a distinct, youngest detrital zircon cluster of at least two zircon that overlap at the 1σ level or of three or more zircons at the 2σ level (e.g. Dickinson and Gehrels, 2009; Gehrels, 2014) was determined for each sample (Figure 3.8). Samples GBLD-004, HA-251-D, and Mamuno-1 yielded weighted-mean $^{207}\text{Pb}^{206}\text{Pb}$ ages for the youngest cluster of three or more zircon analyses at the 2σ level of 1066.4 ± 9.4 Ma (at 95% confidence level, MSWD = 0.88), 1063 ± 11 Ma (MSWD = 0.056), and 1056.0 ± 9.9 Ma (MSWD = 0.68), respectively (Figure 3.8). Sample EISDD-008 yielded a weighted-mean $^{207}\text{Pb}^{206}\text{Pb}$ age of the youngest two analyses that overlap at the 1σ level of 1035 ± 50 Ma (MSWD = 1.4). The 1035 ± 50 Ma age from EISDD-008 is consistent with and within error of the ~1060 – 1050 Ma age group and may indicate a slightly younger, albeit less robust, maximum depositional age for the middle-upper Ghanzi Group (D’Kar and Mamuno formations; Figure 3.8). The combined weighted-mean age of all of the youngest Ghanzi Group detrital zircon grains (n = 12) is 1056.5 ± 8.1 Ma (MSWD = 1.9).

### 3.6.3 Detrital zircon Lu-Hf results

The Lu-Hf results for all individual samples and all combined samples are shown in a plot of initial Hf ($\varepsilon$Hf$_i$) versus U-Pb age (Figures 3.6 and 3.7, respectively). The hafnium values at the time of crystallization ($\varepsilon$Hf$_t$) for the older ~2150 to ~2000 Ma zircon cluster (Group 1) range from -6.5 to +5.5 (Figure 3.7). The Group 2a (~1950 to ~1860 Ma) age cluster has $\varepsilon$Hf$_i$ values that range from -11.7 to +6.6 with much of the data having subchondritic values. The younger ~1820 to ~1700 Ma cluster (Group 2b) is observed in the two western Ghanzi basin samples and one analysis from Mamuno-1 are characterized by $\varepsilon$Hf$_i$ values between -3.1 and +5.5. The one ~1600 Ma zircon and the two ~1500 Ma zircon (Group 3) have distinctly different $\varepsilon$Hf$_i$ values, the former having a negative $\varepsilon$Hf$_i$ value (~4) indicating derivation through melting of older continental crust ($T^{CDM}$ crustal residence age of ~2400 Ma) and the latter having positive $\varepsilon$Hf$_i$ values (+7 to +11) that plot within 5 $\varepsilon$Hf$_i$ values of the depleted mantle curve (Figure 3.7).

The Group 4 age population displays distinct trends in the $\varepsilon$Hf$_i$ plot that coincide with the different subsidiary age populations observed in the relative probability plots. The majority of ~1425 to ~1250 Ma zircon that fall within the subsidiary age population at ~1300 Ma (Group 4a) have $\varepsilon$Hf$_i$ values
between -0.5 and +9.5, although a smaller proportion of the age population has values spreading from -17 to -3 (Figure 3.7). The majority of the ~1250 to ~1180 Ma zircon that fall within the range of the ~1200 Ma population (Group 4b) have εHf\(_t\) values ranging from -11 to +10, with most plotting within ±6 εHf\(_t\) values of the CHUR curve (Figure 3.7). The bulk of ~1180 to ~1080 Ma zircon that fall within the range of the ~1110 Ma population (Group 4c) has εHf\(_t\) values that range from -5.8 to +14.9 (one grain has an εHf\(_t\) value of -14.9). The ~1070 to ~1030 Ma zircon that fall within the range of the ~1060 to ~1050 Ma population (Group 4d) all have positive εHf\(_t\) values between +1.5 to +11 (Figure 3.7).

3.7 Discussion

Radiometric ages determined from the LA-ICPMS U-Pb zircon geochronology are discussed in relation to the various rock formations investigated. These data were used to place maximum age constraints on the timing of basin evolution of the Ghanzi basin and correlate the rocks to different parts...
of the Koras-Sinclair-Ghanzi rift system. The results of the detrital zircon U-Pb and Lu-Hf
geochronology was used to determine the provenance of the Ghanzi Group metasedimentary rocks.

3.7.1 Implications for depositional ages and regional stratigraphic correlations

The multi-grain LA-ICPMS U-Pb zircon age of 1085.5 ± 4.5 Ma reported here for the
Makgabana Hills porphyritic rhyolite flow is ~20 Ma younger than the 1106.1 ± 2 Ma U-Pb zircon age
obtained for a rhyolite from the nearby Mabeleapodi Hills (Figure 3.2; Schwartz et al., 1996) and is
within error of the 1094 ± 20 Ma U-Pb zircon age reported for the Oorlogsende porphyry exposed on the
northern margin of the Ghanzi-Chobe Zone in Namibia (Hegenberger and Burger, 1985). In terms of
regional stratigraphic correlations, the 1106 ± 2 Ma U-Pb zircon age obtained from the Mabeleapodi Hills
(Schwartz et al., 1996) is indistinguishable in age from rhyolites of the Langberg Formation in the
Rehoboth Subprovince (see Figure 3.3; van Schijndel et al., 2014) while the 1085.5 ± 4.5 Ma U-Pb age
obtained from the Makgabana Hills is correlative in age with rhyolites intercalated with metasedimentary
rocks of the Skumok Formation, which overlies the Langberg Formation (Figure 3.3; Hanson et al.,
2006).

The 1085.5 ± 4.5 Ma zircon U-Pb age presents a refined maximum depositional age for the
unconformably overlying Ghanzi Group. Kasbohm et al. (2015) utilized detrital zircon and
paleomagnetic data to correlate the timing of deposition for redbed sequences (Kalkpunt, Aubures, and
Langberg formations within the Kaaien Terrane, Konkiep Subprovince, and Rehoboth Basement Inlier,
respectively; Figure 3.1) within the Koras-Sinclair-Ghanzi rift system at ~1090 Ma. The new Makgabana
Hills rhyolite U-Pb zircon age confirms the assertion of Kasbohm et al. (2015) that the Ghanzi and
Tsumis groups post-date ~1112 to ~1085 Ma magmatism within the Koras-Sinclair-Ghanzi rift system of
Borg (1988). Identification of a youngest zircon age cluster with a statistically reliable (n = 12),
concordant U-Pb age of 1056 ± 9 Ma provides a new maximum depositional age for the middle Ghanzi
Group (including the uppermost Ngwako Pan Formation and cupriferous D'Kar Formation) and indicates
that these formations were at the very least deposited after continent-continent collision with Laurentia
during the amalgamation of Rodinia. Accumulation of the Kuke Formation and lower portions of the
Ngwako Pan Formation is therefore constrained to between ~1085 Ma and ~1060 to ~1050 Ma,
corresponding to the time period on the apparent polar wonder path of the Kalahari Craton as it
approached Laurentia (e.g. Swanson-Hysell et al., 2015).

Minimum depositional age constraints are provided by ~1000 to ~900 Ma Re-Os ages of base
metal sulfides contained within the D’Kar Formation (see Chapter 4). These age data include a Re-Os
molybdenite age of 981 ± 2 Ma and a low-level highly radiogenic Re-Os chalcopyrite age of 914 ± 4
(Hall, 2017); the former is interpreted as the minimum age of deposition of the D’Kar Formation host
rocks. Authigenic xenotime grains from the Ghanzi basin also yielded U-Th-Pb ages ranging from ~925
to ~955 Ma, further strengthening the 981 Ma depositional age of the rocks (Hall, 2017). The combined geochronological data from the Ghanzi basin and other parts of the Koras-Sinclair, Ghanzi rift system indicates that extension within northwest Botswana rift occurred during the amalgamation of Rodinia from ~1050 to ~980 Ma. This combined geological data suggests that the northwest Botswana rift developed into a rifted passive margin setting. This contrasts with the younger Neoproterozoic (<880 Ma) sedimentary successions on the Congo Craton that host world-class sedimentary rock-hosted copper deposits. Those rocks were deposited in an intracontinental rift setting during the break-up of Rodinia. The age of the cupriferous Ghanzi Group suggests that volcano-sedimentary basins of similar age and tectonic setting could host significant sedimentary rock-hosted base metal resources.

3.7.2 Provenance of the Ghanzi Group

To assess plausible provenance source(s) of the Ghanzi Group, the detrital zircon U-Pb and Lu-Hf data were compared to an extensive compilation of available literature on magmatic zircon U-Pb ages and Lu-Hf isotopic data (TDM and TCDM model ages) as well as whole-rock Sm-Nd isotopic data (TDM model ages) from Paleoproterozoic and Mesoproterozoic terranes of the Kalahari Craton (Figures 3.9 and 3.10; see the Supplemental Electronic File for descriptions of terranes, rock types analyzed, U-Pb, Lu-Hf, and Sm-Nd isotopic data, model ages, and complete reference list). The provenance is discussed in terms of major tectonic events that affected the interior and margins of the Kalahari Craton (Figure 3.11).

Possible provenance terranes within the craton interior include the >3.0 Ga Kaapvaal, Zimbabwe, and Grunehogna cratonic components, the ~3.0 to 2.6 Ga Limpopo Belt, and the 2.15 – 2.0 Ga Magondi Belt and Okwa Block (Figures 3.9 and 3.10). The few 2920 – 2460 Ma zircon grains may have been derived from the Limpopo Belt where ~2600 Ma magmatism marks the suturing of the Kaapvaal and Zimbabwe cratons (Xie et al., 2017). The ~2150 to ~2000 Ma U-Pb zircon age group 1 closely matches the timing of events within the footwall terranes of the Magondi Belt and Okwa Block. These rocks have a maximum sedimentation age of 2125 ± 6 Ma (with detrital zircon up to ~3.0 Ga in age) and record widespread plate margin plutonism and volcanism from ~2060 to ~2040 Ga that coincides with the timing of intraplate magmatism in the Bushveld and related igneous complexes (Figures 3.9 to 3.11; Munyanyiwa and Kroner, 1995; Majaule et al., 2001; Mapeo et al., 2001; McCourt et al., 2001; Mapeo et al., 2006). Magmatic activity was followed by granulite facies metamorphism and syn-tectonic magmatism between ~2020 to ~1930 Ma which also affected portions of the Limpopo Belt (Xie et al., 2017). The ~2050 Ma Kubu Island Granite (Magondi Belt) in Botswana contained a 2673 Ma xenocryst and Sm-Nd isotopic data that yielded a TDM model age of 2682 Ma (Figure 3.10; Majaule et al., 2001); this falls within the range of the ~2800 to ~2200 Ga Hf TCDM crustal residence model ages obtained from
Figure 3.9 Precambrian components of the Kalahari Craton within a Gondwana reconstruction highlighting the location of the Meso- to Neoproterozoic northwest Botswana rift (yellow). Modified after Singletary et al. (2003); Jacobs et al. (2008); and de Kock et al. (2014). Abbreviations: BRSZ = Buffels River Shear Zone; CK = Choma- Kalomo Block; ELHSZ = Excelsior-Lord Hill Shear Zone; HRTB = Hartbees River Thrust; Kw = Kwando Complex; MA = Matchless Amphibolite; MF = Molopo Farms Complex; MSZ = Mwembeshi Shear Zone; OB = Okwa Block; PTVSZ = Pofadder-Tantalite Valley Shear Zone; RBI = Rehoboth Basement Inlier; RC = Roibok Complex; SDTZ = Southern Damara Thrust Zone; SFZ = Southern Front Zone; TT = Tugela Thrust; XC = Xade Complex.
Figure 3.10 Comparative chart displaying the Ghanzi Group detrital zircon U-Pb and Lu-Hf isotopic data and U-Pb and radiometric isotopic data (crustal residence ages) for the ~1.0 Ga Kalahari Craton. See the Supplemental Electronic for detailed U-Pb zircon age data, Lu-Hf zircon T^CDM model ages, and whole-rock Sm-Nd T^DM model ages with complete references.
Figure 3.11 Terrane correlations in the Kalahari Craton. See text for detailed explanations and references.
the group 1 detrital zircon. The sparse number of Archean-aged zircon (3 zircon), which could represent recycled zircons from the Magondi Belt, and the small relative proportion of the ~2150 to ~2000 group 1 zircon population (15 zircon) suggests that the cratonic interior was a minor sediment source for the Ghanzi Group sedimentary rocks analyzed in this study.

The ~1950 to ~1860 Ma U-Pb zircon group 2a has εHf values between -11 and +7 and TCDM model ages between 3000 and 2200 Ma; few of the detrital zircon have positive εHf values with more juvenile TCDM model ages approaching ~2050 Ma. The detrital zircon ages and Lu-Hf data correspond well to age and isotopic data for ~1860 to ~1920 Ma bimodal magmatic events recorded around the peripheries of the Rehoboth Subprovince, a deeply buried, aeromagnetically distinct Paleoproterozoic shield with suspected Archean roots that is only exposed in the Kheis Belt, the Rehoboth Basement Inlier, the Richtersveld Subprovince (Figures 3.9 and 3.10).

The ~1915 Ma Hartley Formation basalts (Kheis Belt, Figures 3.9 and 3.10) have TDM model ages of ~2650 to ~2450 Ma (Cornell et al., 2016) that are similar to some of the older group 2a zircon (Figure 3.9). However, most of the detrital zircons have slightly younger ages between ~1900 and ~1865 Ma. The ~1870 Ma Elim Formation (Rehoboth Basement Inlier; Figure 3.9) contains suprachondritic εHf values of +4.2 to +6.3 (Figure 3.10; van Schijndel et al., 2014) and may be a source for the few detrital zircons with more juvenile Hf isotopic signatures.

The ~1.9 Ga Richtersveld Magmatic Arc records volcanism (Orange River Group) and magmatism (Vioolsdrif Suite) between ~1910 to ~1865 Ma within the Richtersveld Subprovince (Figures 3.9 to 3.11; Macey et al., 2017). Calculated Sm-Nd whole-rock TDM model ages of ~2400 to ~2150 Ma (and up to ~2800 Ma in one sample; Figure 3.9; Reid, 1997; Macey et al., 2017) are in broad agreement with the Hf TCDM model ages obtained from the detrital zircon population (Figure 3.10). These data suggest that the Paleoproterozoic portions of the Rehoboth Subprovince may have contributed directly to the sediment supplied into the Ghanzi basin.

Igneous rocks with ages similar to the ~1860 to ~1700 Ma U-Pb zircon age group 2b have only been reported from the Rehoboth Basement Inlier (Figure 3.9; van Schijndel et al., 2014). These rocks include felsic to intermediate volcanic rocks of the Kalkbrak, Gaub Valley, and Marienhof formations with ages of ~1826 Ma, ~1780-1750 Ma, and ~1770-1750 Ma, respectively, and their plutonic equivalents, the ~1820-1810 Ma Kangas Metamorphic Complex, the ~1780 Ma Piksteel Granite, the ~1764-1740 Ma Weener Igneous Complex, the ~1720 Ma Mt. Barry granodiorite, and the ~1755 Ma Brack amphibolite located in the Hohewarte Complex (Ziegler and Stoessel, 1993; Becker et al., 1996; Nagel et al., 1996; Mapani et al., 2014; van Schijndel et al., 2014). The late Paleoproterozoic Rehoboth Basement Inlier rocks (Figure 3.99) are defined by predominantly chondritic to suprachondritic εHf values (+0.5 to +5.4), with only parts of the Gaub Valley volcanoclastic rocks recording subchondritic
$\varepsilon$Hf values (-2.7 to -0.5; van Schijndel et al., 2014). These values are strikingly similar to those obtained for the group 2b detrital zircon (Figure 3.10), suggesting that the Rehoboth Basement Inlier is the likely source of these zircon.

The ~1608 Ma zircon (one zircon in group 3) most closely resembles the 1604 ± 33 Ma U-Pb zircon age obtained from a volcanic tuff in the Palapye Group, located in eastern Botswana (Figure 3.10; Mapeo et al., 2004). The tuff also contained xenocrysts with U-Pb ages of ~2600 and ~2035 Ma, which correlate with the ~2450 Ma $T_{DM}^{C}$ model age for the detrital zircon (Figure 3.10). In contrast, the two ~1500 Ma group 3 zircon have suprachondritic $\varepsilon$Hf values, one of which plotted close to the depleted mantle curve. Several ~1600 to ~1500 Ma xenocrysts, one with a reported $\varepsilon$Hf value of +3.9, have been reported from ~1230 Ma intrusions in the Konkiep Subprovince (Figures 3.9 to 3.11; Cornell et al., 2015) and the Grünau area of the Gordonia Subprovince (Figures 3.9 to 3.11; Cornell and Pettersson, 2007, Bial et al., 2015), as well as from a ~1037 Ma intrusions in the Bushmanland Subprovince (Figures 3.9 to 3.11; Robb et al., 1999). The Namaqua Sector could be a possible source for the ~1500 Ma group 3 zircon.

The ~1420 to ~1000 Ma U-Pb zircon age group 4 could have been sourced from a number of Mesoproterozoic terranes on the margins of the Kalahari Craton. The eastern edge of the Kalahari Craton was a convergent margin from ~1200 to ~1050 Ga, indicated by terranes of the Natal Sector, the Maud Belt in present day Antarctica, and a Mesoproterozoic portion of the Mozambique belt, the Nampula Complex (Figure 3.9; Hanson et al., 2004; Jacobs et al., 2008). The Pan-African (~570 to ~530 Ma) Lurio Belt represents a suture between the Nampula Complex and slightly younger magmatic rocks (1020 to 950 Ma) of the southern Irumide Belt, Unango Complex, and Marupa Complex (Figure 3.9), which are presumed to have formed as an active margin on the Congo-Tanzania Craton (Bingen et al., 2009).

The southeastern craton margin contains ~1200 to ~1100 Ma back-arc and island arc terranes that were successively accreted to the southeastern margin of the Kalahari Craton by 1000 Ma (Figure 3.11; Jacobs et al., 2008). These juvenile island arc terranes have primarily suprachondritic $\varepsilon$Hf and $\varepsilon$Nd values with corresponding Sm-Nd $T_{DM}$ model ages between ~1700 to 1100 Ma (Figure 3.10; Grantham et al., 2011). The northeastern margin is characterized by continental arc volcanism within portions of the Grunehogna Craton and Mozambique Belt (Marschall et al., 2013). Lu-Hf $T_{DM}^{C}$ and Sm-Nd $T_{DM}$ model ages obtained from the continental arc plutonic and volcanic rocks from Grunehogna and Mozambique indicate the involvement of ~3805 to ~3450 Ma (not shown) and ~3300 to ~3000 Ma crust, respectively (Figure 3.10; Manhica et al., 2001; Marschall et al., 2010; Grantham et al., 2011).

Although the U-Pb ages from the Natal-Maud-Mozambique belts have a moderate correlation to the U-Pb zircon age groups 4b-d, the predominantly juvenile isotopic character of the majority of these rocks contrasts with the mixed juvenile and crustal reservoir isotopic character of the Ghanzi Group.
detrital zircons. Additionally, the relative lack of Archean to Paleoproterozoic zircon supplied from the craton interior suggests that the younger Mesoproterozoic terranes that surround the southeastern and northeastern margins of the Kalahari Craton were not likely source terranes for the Ghanzi Group. This interpretation is supported by the presence of sedimentary back-arc basins (e.g. Mfongosi Group, Tugela Terrane, Natal Sector; Basson et al., 2004) and late tectonic extensional basins (e.g. Alto Benifica Group, Nampula Complex, Mozambique Belt; Thomas et al., 2010) adjacent to and/within these belts.

Magmatic rocks that could have supplied at least part of the group 4 zircon population occur on the northern/northwestern margins of the northwest Botswana rift. The ~1425 to ~1250 Ma U-Pb zircon age group 4a display εHf values that plot between depleted mantle and chondritic values corresponding to TCDM model ages of ~2050 to ~1450 Ma, although some zircon display distinctively enriched εHf values between -5 and -17 with TCDM model ages up to ~2900 Ma. Hanson et al. (1988) and Bulambo et al., (2006) reported ~1370 to ~1285 Ma U-Pb zircon ages from a large plutonic complex that comprises much of the Choma-Kalomo Block in southern Zambia (Figures 3.9 to 3.11). A single Sm-Nd TDM model age from the Zongwe orthogneiss indicates a crustal residence age of >2000 Ma (Figure 3.10; Hanson et al., 1988). Kampunzu et al. (2000) suggested that a ~1363 Ma detrital zircon population obtained from the Ghanzi Group in northeastern Botswana (Goha Hills) may have been derived from the Choma-Kalomo Block. However, there is still strong debate as to whether the Choma-Kalomo Block formed part of the Kalahari Craton at the time (Glynn et al., 2014) or if it represents an exotic block that rifted off of the Congo craton and was subsequently caught up between the two cratons during Pan-African orogenesis (Bulambo et al., 2006).

To the west of the Goha Hills area, the northwest Botswana rift is bounded to the northwest by ~1200 to ~1150 Ma granitic gneisses of the Kwando Complex (Figure 3.9). These rocks were accreted to the Kalahari margin prior to intrusion of a ~1110 Ma gabbro to diorite body (Figure 3.11; Singletary et al., 2003). A whole-rock Sm-Nd TDM model age of ~1295 Ma for the granitic gneiss indicates a relatively short crustal residence period (Figure 3.10; Singletary et al., 2003). Although the scarce data from the Kwando Complex may reflect some of the more juvenile group 4b zircon, the Ghanzi Group Lu-Hf data overwhelmingly indicates source terranes that are characterized by reworking of older Paleoproterozoic crust.

The Namaqua Sector records the 600-million-year evolution of the Namaqua Wilson Cycle that involved several highly contrasting terranes and several different episodes of magmatism, terrane amalgamation, and final continental collision by ~1060 to ~1000 Ma (Figure 3.11; Miller, 2012). Initial fragmentation of the Rehoboth Subprovince into several distinct blocks began at ~1675 Ma, indicated by the maximum depositional age of the Billstein Formation in the Rehoboth Subprovince (van Schijndel et al., 2014). The rift to drift transition is indicated by ~1750 to ~1570 Ma Lu-Hf and Sm-Nd model ages for...
the ~1700 to ~1100 Ma Koras Group, ~1600 to ~1501 Ma zircon core and xenocrysts from the Bushmanland Terrane, and ~1500 Ma zircon cores from the Gordonia Subprovince (Figures 3.9 to 3.11). The initiation of subduction-related magmatism is marked by ~1470 Ma Kairab Formation (Konkiep Subprovince; Hoal, 1990). This was followed by emplacement of ~1380 to ~1330 Ma active margin and back-arc granitoids within the Konkiep Subprovince and the Kaairen Terrane (Figure 3.11; Hoal and Heaman, 1995; Pettersson et al., 2007; Cornell et al., 2015). The Konkiep Subprovince granitoids display Lu-Hf $T^{CDM}$ model ages between ~1700 to ~1450 Ma while the Kaairen Terrane granitoids display Sm-Nd $T^{DM}$ model ages between ~2300 to ~1500 Ma (Figure 3.10; Cornell et al., 2015; Pettersson, 2008).

Accretion of the Konkiep Subprovince along the northwesterly trending Namaqua Front occurred from ~1330 to ~1300 Ma (Figure 3.11; Miller, 2012). The subsequent subduction-polarity reversal initiated convergent active margin and back-arc volcanism within these terranes from ~1300 to ~1260 Ma. These magmatic rocks have Lu-Hf $T^{CDM}$ and Sm-Nd $T^{DM}$ model ages between ~2300 and ~1500 Ma that indicate mixing between a juvenile mantle source and older Paleoproterozoic crust (Figure 3.10; Bailie et al., 2012, Cornell et al., 2015). Subduction along the western margins of the newly accreted terranes initiated the development of the Areachap Terrane, an oceanic arc terrane that developed between ~1290 and ~1240 Ma that contains predominantly juvenile, MORB-like Sm-Nd isotopic signatures and $T^{DM}$ model ages between ~1750 to ~1330 Ma (Figures 3.10 to 3.11; Cornell et al., 1986; Bailie, 2008; Pettersson, 2008; Bailie et al., 2010). The broad range of $T^{CDM}$ and $T^{DM}$ model ages obtained from these terranes are in excellent agreement with the Lu-Hf $T^{CDM}$ model age data for the group 4a Ghanzi Group detrital zircon population (Figure 3.10).

The majority of the ~1230 to ~1150 Ma U-Pb zircon age group 4b display trends towards chondritic to sub-chondritic $\varepsilon$Hf, $t$ values ($T^{CDM}$ model ages between ~2500 and ~1500 Ma) that suggest mixing/contamination by and/or recycling of older Paleoproterozoic to Archean crustal material. This age group broadly reflects the second phase of the Namaqua Wilson Cycle that involved the development of the ~1230 to ~1200 Ma Rehoboth Magmatic Arc (Figure 3.11). The Gamsberg Granite Suite and Nückopf Formation rhyolites (Rehoboth Subprovince), as well as hybrid intrusions within the Konkiep Subprovince display chondritic to sub-chondritic $\varepsilon$Hf, $t$ values between -4.1 and +2 corresponding to $T^{DM}$ model ages between 2200 and 1850 Ma (Figure 3.10; van Schijndel et al., 2014). A mantle-like $\delta^{18}$O$_{zrc}$ value of 4.99‰ obtained from the Gamsberg Granite Suite suggest derivation from older Paleoproterozoic mafic lower-crust (van Schijndel et al., 2014) while Lu-Hf $T^{DM}$ model ages between ~1900 and ~2200 Ma and zircon cores with U-Pb ages between ~1755 and ~1260 Ma obtained from the Konkiep Subprovince point towards mixing between a juvenile mantle component and older crustal material in an active margin setting (Figure 3.10; Cornell et al., 2015).
Cessation of subduction throughout the Namaqua Sector occurred at ~1200 Ma due to amalgamation of the Gordonia Subprovince and Kaaien and Areachap terranes to the southwestern margin of the Rehoboth Subprovince, deformation in the Kheis Belt, and finally docking of the Bushmanland Subprovince and Garies Terrane with the southern margin of the Richtersveld Subprovince from ~1200 to ~1103 Ma (Figure 3.11; Miller, 2012, and reference therein). This is referred to as the ~1200 Ma O’okiepian episode of Namaquan orogenesis (Figure 3.11; Clifford et al., 2004; Miller, 2012) that involved voluminous syn-tectonic granitoid emplacement within the Konkiep Subprovince and the Areachap terrane (early phases of the Keimoes Intrusive Suite). These granitoids have Lu-Hf T<sub>CDM</sub> and Sm-Nd T<sub>DM</sub> model ages primarily between ~2200 and ~1850 Ma; few model ages up to ~2780 Ma point towards recycling of older Archean continental crust (Figure 3.10; Cornell et al., 2015; Nethenzheni, 2016). In contrast, the ~1200 to ~1150 Ma syn-tectonic granite gneisses within the Gordonia Subprovince and the Bushmanland and Garies terranes have Sm-Nd T<sub>DM</sub> model ages between ~2300 and ~1360 Ma suggesting involvement of both juvenile mantle material and reworking of older Paleoproterozoic crust (Figure 3.10; Yuhara et al., 2001; Pettersson, 2008; Samskog, 2009 and Nordin, 2009, unpublished data in Miller, 2012). In addition to syn-tectonic magmatism, decompression melting of subduction-related subcontinental lithospheric mantle following slab detachment led to the A-type bimodal volcanism within the Kaaien Terrane as indicated by predominantly negative εNd values obtained from the ~1170 Ma lower Koras Group (Figures 3.10 to 3.11; Bailie et al., 2012).

Most of the 1120-1080 Ma U-Pb zircon age group 4c have suprachondritic εHf<sub>t</sub> values between -1.2 and +15; one grain from Mamuno-1 records a εHf<sub>t</sub> value of -15 and a T<sub>CDM</sub> model age of ~2600 Ma (Figure 3.7). This age range reflects a period widespread bimodal plate margin and mafic intraplate magmatism throughout the Kalahari craton related to the Umkondo LIP (Figure 3.9; Hanson et al., 2004). Right-lateral shearing throughout the Namaqua Sector resulted in emplacement of ~1112 to ~1080 Ma bimodal volcanic rocks of the Koras Group and upper Nazerus Group within extensional basins (Borg, 1988). Coeval syn-tectonic granitoids of the Keimoes and Gamsberg granite suites were emplaced in all the Namaquan terranes except for the Bushmanland and Richtersveld subprovinces (Figure 3.11; Miller, 2012, and references therein). Lu-Hf and Sm-Nd isotopic data from the ~1120 to ~1080 Ma Namaquan magmatic rocks indicate a wide range of near-depleted mantle values to chondritic values with T<sub>CDM</sub> and T<sub>DM</sub> model ages ranging from ~3350 to 1200 Ma (Pettersson, 2008; Pettersson et al., 2009; Bailie et al., 2011b; Nethenzheni, 2016); these data match well with the Ghanzi Group Lu-Hf isotopic data for the U-Pb zircon age group 4c (Figure 3.10). Some of the more enriched group 4c zircon could have been sourced from extensive mafic intrusive complexes (e.g. Xade, Tsetseng, and Tshane complexes in Botswana), dyke swarms, and sills of continental tholeiitic composition that were rapidly emplaced into
all the major components of the Kalahari craton during the 1112 to 1106 Ma Umkondo LIP event (Figure 3.9 to 3.10; Hanson et al., 2004; de Kock et al., 2014 and references therein).

Paleomagnetic polar wander paths for the Kalahari Craton indicate an abrupt change in plate motion at approximately 1100 Ma, setting the Kalahari Craton on a collision path with Laurentia (Hanson et al., 2004; Li et al., 2008; Swanson-Hysell et al., 2015). The Bushmanland and Garies terranes were subjected to intense deformation, recumbent folding, and thrusting between ~1110 and ~1070 Ga as Kalahari approached Laurentia. The subsequent collisional events from ~1060 to ~1020 Ga, referred to as the Klondikean episode of Namaquan orogenesis by Clifford et al. (2004; Figure 3.11), resulted in upright folding accompanied by intrusions of the Spektakel and Koperberg Suites, granulite facies metamorphism, and final pegmatite emplacement at ~1000 Ma (Miller, 2012, and references therein). Minor magmatism accompanied up to 12 km of uplift within the high-grade Gordonia Subprovince; xenocrysts ranging from ~1600 to ~1135 Ma indicate contamination from the surrounding country rocks (Robb et al., 1999). Minor late tectonic intrusions of the Gamsberg Granite Suite and the Seeis granite gneiss occurred at ~1065 to ~1020 Ma within the Rehoboth Subprovince (Miller, 2008; Mapani et al., 2014). Scarce Lu-Hf isotopic data from the Seeis granite yielded a Lu-Hf $T^{CDM}$ model age between ~1630 and ~1580 Ma (Mapani et al, 2014). The data from the Klondikean magmatic rocks are similar to that of the ~1060 to ~1050 Ma U-Pb zircon age group 4d obtained from the Ghanzi Group (Figure 3.10).

The U-Pb and Lu-Hf isotopic data obtained from the Ghanzi Group metasedimentary rocks strongly indicates provenance from a terrane that contained both Paleoproterozoic and Mesoproterozoic crustal components, with much of the latter resulting from reworking of the former. The terranes surrounding the Paleoproterozoic Rehoboth Subprovince, namely the Namaqua Sector and the Rehoboth Basement Inlier, record tectonic and magmatic events that closely match the detrital zircon U-Pb and Lu-Hf signature of the Ghanzi Group metasedimentary rocks and are therefore considered the main provenance sources for the Ghanzi Group and the northwest Botswana rift (Figure 3.10).

Modie (1996) described sedimentary structures and paleo-current directions within the Ghanzi Group that suggested that the predominant sediment transport direction in the basin was to the northeast, along the axis of the basin. These observations, coupled with the presence of Paleoproterozoic detrital zircon grains that were likely derived from the Rehoboth Subprovince, suggests that a major portion of the ~1420 to ~1000 Ma U-Pb zircon age group 4 was probably derived from sources located within and surrounding the periphery of the Rehoboth Subprovince, including the ~1460 to ~1000 Ma Namaqua Sector (Eglington, 2006; Miller, 2012).

The orogeny along the Namaqua-Grenville front likely fed major continental fluvial systems that transported sediment across the interior of the Rehoboth Subprovince that was eventually funneled towards the central and western Ghanzi basin (Figure 3.12). Sediment input from the craton
Figure 3.12 Reconstruction of the Kalahari Craton at ~1.0 Ga. Modified after Jacobs et al. (2008) with location of the northwest Botswana rift (yellow). Black arrows indicate probable major and minor provenance sources for the Ghanzi Group metasedimentary rocks (northwest Botswana rift) based on U-Pb and Lu-Hf isotopic data from detrital zircon. Blue arrows indicate plausible fluvial sedimentary transport paths that drained the ~1.06 – 1.02 Namaquaan-Grenvillian orogeny.

interior/southeastern footwall and terranes forming the northern margins of the rift may have been limited to more proximal sedimentary deposits such as footwall fans, with only minor material reaching the interior of the basin (Figure 3.12). Samples from the western Ghanzi basin display more distinct U-Pb
and Lu-Hf isotopic data that suggest a stronger affinity to more proximal sources in the Rehoboth Subprovince. These new isotopic data indicate that the Ghanzi Group sediments were deposited in an actively evolving rifted continental margin setting on an outboard (oceanward) margin of the craton during the assembly of Rodinia. Rifting of the continental margin was possibly driven by far-field extensional stresses resulting from accretion and indentation tectonics on the opposing cratonic margins between 1100 and 1000 Ma.

The timing of basin formation contrasts with the younger Neoproterozoic (<880 Ma) successions on the Congo Craton that host world-class sedimentary rock-hosted copper deposits, which developed in an intracontinental rift setting during the break-up of Rodinia. The age of the cupriferous Ghanzi Group suggests that volcano-sedimentary basins of similar age and tectonic setting could host significant sedimentary rock-hosted base metal resources, one significant known example of which is the White Pine district in Michigan, USA, where copper deposits are hosted by 1110 to 1100 Ma bimodal volcanic rocks of the Midcontinent rift system and overlying Meso- to Neoproterozoic mudstones of the Nonesuch Formation (e.g. Mauk et al., 1992).

3.8 Conclusions

The LA-ICPMS multi-grain U-Pb zircon age of 1085.5 ± 4.5 Ma obtained from a rhyolite flow of the Makgabana Hills is the youngest igneous rock dated in the northwest Botswana rift. The date sets a new maximum depositional age for the unconformably overlying Ghanzi Group. The rhyolite flows and volcaniclastic rocks of the Makgabana Hills are correlated with the uppermost Nazerus Group in the Rehoboth Subprovince. This new age marks the youngest age reported from the ~1.23 to ~1.09 Ga Rehoboth Magmatic Arc on the northwestern margin of the Kalahari Craton.

The unconformably overlying Ghanzi Group metasedimentary sequence was derived from both Paleoproterozoic and Mesoproterozoic source terranes. Youngest clusters of three or more overlapping zircon from three samples indicate a maximum depositional age of the middle Ghanzi Group, including the uppermost Ngwako Pan Formation, of ~1065 to ~1055 Ma. Minor detrital zircon with ages of ~2150 to ~2000 Ma were likely derived from the adjacent Magondi Mobile Belt based on similarities in U-Pb ages and Lu-Hf and Sm-Nd isotopic data; the Okwa Block contains similar U-Pb ages but lacks Lu-Hf and Sm-Nd isotopic data for correlations. Younger (~1950 to ~1700 Ma) detrital zircon was likely derived from exposed portions of the Paleoproterozoic Rehoboth Subprovince, including the Rehoboth Basement Inlier, Richtersveld Subprovince, and the Kheis Belt.

Hf isotopic data indicate that the majority of the Mesoproterozoic detrital zircon was ultimately derived from extensively reworked Paleoproterozoic crust up to ~2050 Ma in age with some zircon showing evidence for derivation from reworking of early Paleoproterozoic and Neo- to Mesoarchean crust involved in the generation of magmatic melts. The detrital zircon isotopic data from the Ghanzi
Group metasedimentary rocks broadly reflects the ~1600 to ~1000 Ga Wilson-cycle evolution of the Namaqua-Natal-Maud orogenic province along the western, southern, and eastern margins of the Kalahari Craton. Lu-Hf and Sm-Nd isotopic data from the Natal Sector and Dronning Maud Land indicate that ~1200 to ~1060 Ga arc magmatism and accretionary events along the eastern margin of the Kalahari Craton primarily involved a juvenile mantle source, although continental arc magmatism is inferred to have occurred along portions of the Grunehogna Craton and the Nampula Complex in Mozambique. However, sediments derived from these terranes were likely deposited in adjacent flysch and/or intermontane basins rather than being transported across the interior of the craton to the Ghanzi basin.

The detrital zircon U-Pb age and Lu-Hf isotopic data obtained from the Ghanzi Group are consistent with Mesoproterozoic magmatic ages and radiogenic isotopic data (Lu-Hf and Sm-Nd) from the Rehoboth Subprovince and the Namaqua Sector. Isotopic data from these terranes indicate substantial reworking of Paleoproterozoic crust during ~1400 to ~1000 Ma tectono-magmatic events on the western margin of the Kalahari Craton that culminated in continent-continent collision with Laurentia. Detritus derived from the extensive orogenic belt could have been transported across the interior of the Rehoboth Subprovince by continental fluvial systems before entering the Ghanzi basin and being dispersed and deposited through fluvial-deltaic and shallow shelf processes. The revised timing of deposition of the Ghanzi Group presented here indicates that the northwest Botswana rift developed as a rifted margin basin during the assembly of Rodinia, in contrast to other Neoproterozoic basins that developed later during the break-up of Rodinia. These finding can have significant implications for the metallogeny of the Kalahari Copperbelt and other basins of similar age.

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4.1 Abstract

Rhenium-osmium geochronology of molybdenite-(bornite), chalcocite-idaite, and arsenopyrite together with U-Th-Pb laser ablation split-stream inductively coupled mass spectrometry (LASS ICP-MS) analyses of xenotime grains reveal a prolonged history of events within the Ghanzi Ridge area of the Kalahari Copperbelt in northwestern Botswana. A vein-hosted low-level highly-radiogenic (LLHR) chalcocite-idaite separate from a quartz-calcite-chalcocite-idaite-bornite-native silver vein (Zone 6 prospect in the Northeast District of the central Kalahari Copperbelt) and a xenotime inclusion intergrown with vein-hosted chalcopyrite-molybdenite (Northeast Mango Two prospect, Northeast District) yield Re-Os and U-Th-Pb ages of 549.0 ± 11.2 Ma and 538 ± 8 Ma, respectively. The ages pre-date the timing of ~535 to ~530 Ma peak metamorphism in the Damara Belt and likely correspond to northwest to southeast shortening in the Ghanzi-Chobe zone. A fracture-controlled, coarse-grained molybdenite-(bornite) assemblage that overprints structurally-controlled alteration (iron reduction/bleaching) of footwall sedimentary rocks at the Northeast Fold deposit in the Banana Zone district to the southwest of the Northeast District yield a Re-Os age of 515.9 ± 2 Ma. This suggests a post-peak metamorphic mineralizing event in the Kalahari Copperbelt.

The sample containing the 538 ± 8 Ma xenotime inclusion also contains abundant disseminated molybdenite that replaced bedded framboidal pyrite grains and remnant organic matter coatings within the wall rock adjacent to the hydrothermal vein. Veinlet- and cement- to replacement-style molybdenite occurs within a wall rock fragment that was sheared and encased by a quartz-calcite-chalcopyrite vein. The molybdenite yielded a Re-Os age 981 ± 3 Ma. This suggests that the Re-Os systematics of the framboidal pyrite was inherited, wholesale, during replacement by the hydrothermal molybdenite. These data indicate that Re-Os systematics of sulfides can survive *in-situ* on solid state recrystallization to other sulfide species. The 981 ± 3 Ma Re-Os molybdenite age also provides a new minimum depositional age constraint for the D’Kar Formation host rocks. Several authigenic xenotime grains along the margin of the 549.0 ± 11.2 Ma quartz-calcite-chalcocite-idaite-bornite-native silver vein yield three concordant U-Th-Pb ages indicating xenotime growth at ~950 to 925 Ma, whereas several highly discordant analyses indicate that the grains were affected by a subsequent event(s) that resulted in disturbance of the isotopic
system. Ages for xenotime grains further support a depositional age of >950 Ma for the D'Kar Formation. Coarse-grained, prismatic arsenopyrite grains that are over-printed by syn-kinematic quartz-calcite-chalcopyrite-sphalerite-galena cleavage lenticels (Zone 5 deposit, Northeast District) yield separates with marginally LLHR character and Re-Os ages that range from 670 ± 9 to 630 ± 8 Ma. The age range corresponds to the period between the Sturtian and Marinoan glaciations (Cryogenian non-glacial interlude), and a history that is older than the main syn-kinematic, copper-bearing hydrothermal mineralization in the Kalahari Copperbelt associated with Pan-African orogenic events.

4.2 Introduction

Geochronology from many of the world’s sedimentary copper districts shows that mineralization in these systems can include early diagenetic events and events during basin inversion and metamorphism (Hitzman et al., 2010). The Ghanzi-Chobe zone in Botswana and the Southern Foreland zone in Namibia (Figure 4.1) represent the foreland fold-thrust belt of the late Neoproterozoic-Cambrian Damara Orogeny (e.g. Miller, 2008). The Ghanzi-Chobe zone in Botswana affected a late Mesoproterozoic to early Neoproterozoic volcano-sedimentary basin termed the northwest Botswana rift (Key and Ayres, 2000). The region was coined the Kalahari Copperbelt by Borg and Maiden (1986a) due to the widespread occurrence of sedimentary rock-hosted copper-silver deposits and prospects (Figure 4.1). Deposits generally occur along a regional litho-chemical redox trap that is characterized by shallow marine siliciclastic and carbonate rocks that disconformably overlie ‘red bed’ continental clastic rocks (Figure 4.1; Schwartz et al., 1995). Economically important accumulations of copper sulfides occur in structurally-controlled quartz-carbonate veins and as disseminations throughout the host rocks. The Cu-Ag mineralization in the Kalahari Copperbelt has been attributed to syngenetic and/or early and late diagenetic processes. Diagenetic mineralizing processes were inferred from the textures of sulfides that were intergrown with diagenetic mineral phases and sulfur stable isotopes indicative of bacterial reduction of sulfate (Martin, 1965; Baldock et al., 1976; Ruxton, 1986; Ruxton and Clemmey, 1986; Schwartz et al., 1995; Borg, 1995). Epigenetic-only processes were inferred based on the predominance of cross-cutting quartz-calcite veins that contain an identical vertical zonation of copper sulfide minerals to that of disseminated sulfides (Sillitoe et al., 2010; Maiden and Borg, 2011). Multi-stage processes involving overprinting and/or remobilization of diagenetic precursor or low-grade base metal sulfides during epigenetic mineralizing processes have also been invoked (Borg and Maiden, 1986b, 1987, 1989; Borg, 1988, 1995; Walraven and Borg, 1992; Van der Heever et al., 2009; Gorman, 2013, Hall, 2013a).

Prior to the now widespread use of dating techniques such as Re-Os sulfide and U-Th-Pb monazite/xenotime, the only reported ages for mineralization in the Kalahari Copperbelt came from whole-rock Pb isotopic signatures of mineralized and non-mineralized ‘laminites’ from the Klein Aub
Figure 4.1 Bedrock geologic map and Cu-Ag deposits of the Kalahari Copperbelt in Botswana. Geology is inferred from outcrop, drill core, and regional aeromagnetic datasets. MH = Mabeleapodi Hills; NH = Nguneakau Hills; KH = Kgwebe Hills; MkH = Makgabana Hills; inset: GCz = Ghanzi-Chobe zone; KCB = Kalahari Copperbelt (white outline in inset); NB = Namaqua Belt; SFz = Southern Foreland zone of the Damara Orogen. Deposits and prospects with Re-Os and U-Th-Pb analyses presented here are printed in bold italic font and underlined in the key. Bold print with asterisks (*) indicate deposits with previously reported Re-Os ages from Hall (2013). Inset map of the regional Precambrian basement rocks in southern Africa with the location of the Ghanzi-Chobe zone and the Ghanzi Ridge area, the focus of this study.

deposit in Namibia suggesting a ~600 Ma mineralizing event that overprinted an earlier period of mineralization (Walraven and Borg, 1992). The Pb isotope systematics also indicate post-mineralization fluid-rock interaction (Walraven and Borg, 1992).
The predominantly bedding-parallel hydrothermal vein systems that characterize deposits of the Kalahari Copperbelt have been attributed to flexural-slip processes that operated during progressive Damara (~600 to 480 Ma) deformation in the Ghanzi-Chobe zone (Davies, 2013; Hall, 2013a). However, recent Re-Os geochronology of vein-hosted chalcopyrite and bornite separates from the Boseto district yielded ages of 1012 ±17, 914 ± 4, and 469 ± 27 Ma (Hall, 2013a), inviting a larger study of the timing of hydrothermal events in the Kalahari Copperbelt and the age of the host rocks as well. This study presents new petrographic data coupled with Re-Os geochronology (molybdenite-bornite, chalcocite-idaite, and arsenopyrite) and U-Th-Pb LA-ICPMS analyses of xenotime from several deposits and prospects within the Ghanzi Ridge portion of the Kalahari Copperbelt to provide a more comprehensive history of depositional, diagenetic, metamorphic and mineralizing events. The results suggest that diagenetic sulfides, in-situ organic matter, and mobile hydrocarbons may also have been involved in an apparently complex, multi-stage series of mineralization events stretching from diagenesis to basin inversion.

4.3 Geologic background

The Kalahari Copperbelt in northwest Botswana is situated on the northwestern margin of the Kalahari (Kaapvaal-Zimbabwe) craton in southern Africa (Figure 4.1). It contains several sedimentary rock-hosted Cu-Ag deposits and prospects around the Ghanzi Ridge (Figure 4.1), a prominent structural ridge defined by isolated basement inliers within the Kalahari Desert. The deformed Precambrian basement rocks at the heart of the Ghanzi Ridge form part of a ~600-km-long, geophysically defined northeast-southwest trending fold and thrust belt termed the Ghanzi-Chobe zone that extends from the northeastern corner of Botswana to the western border with Namibia (Figure 4.1, inset; Modie, 1996; Lehmann et al., 2015). This fold and thrust belt deformed late Meso- to early Neoproterozoic volcano-sedimentary rocks, the Kgwebe Formation and Ghanzi Group, that were deposited in the northwest Botswana rift (Key and Ayers, 2000).

4.3.1 Host rocks

Deposition of the Kgwebe Formation is inferred to represent the initial phase of intracratonic rifting along the northeast-trending northwest Botswana rift (Modie, 1996; Key and Ayers, 2000). It consists of a bimodal volcanic sequence of rhyolite, basalt, minor dacite, and intercalated metasedimentary rocks (Schwartz et al., 1995; Kampunzu et al., 1998). The Kgwebe Formation has been dated at several localities through various techniques. In the Chobe basin/zone (northeastern Botswana), dates include a porphyritic rhyolite with a 981 ± 43 Ma Rb-Sr whole-rock errorchron age (Key and Rundle, 1981) as well as ID-TIMS zircon weighted-mean \(^{207}\text{Pb}/^{206}\text{Pb}\) ages of 1107.5 ± 0.5 Ma (granite), 1107.4 ± 2.1 Ma (granite), 1107.0 ± 0.8 Ma (gabbro), and 1106.2 ± 3.6 Ma (rhyolite; Singletary et al., 2003). Based on the ID-TIMS zircon age of 1106.2 ± 3.6 Ma from nearby rhyolites (Singletary et al.,
2003), the Rb-Sr whole rock errrorchron age (Key and Rundle, 1981) is reinterpreted here to reflect a post-
crystallization thermal event that affected the Kgwebe Formation rocks, such as Rb-Sr isotopic resetting
by metamorphic fluids in shear zones (e.g. Hickman and Glassley, 1984) or hydration during uplift (e.g.
Evans et al., 1995). In the Ghanzi basin/zone, the Mabeleapodi Hills rhyolite yielded an ID-TIMS
\(^{207}\) Pb/\(^{206}\) Pb age of 1106 ± 2 Ma (Schwartz et al., 1996) while LA-ICPMS geochronology of the nearby
Makgabana Hills rhyolite yielded a weighted-mean \(^{206}\) Pb/\(^{238}\) U age of 1085.5 ± 4.5 Ma (see Chapter 3). In
the northwestern Ghanzi basin/zone, the Oorlogsende porphyritic rhyolite yielded and ID-TIMS age of
1092 ±18 Ma (Hegenberger and Burger, 1985).

The unconformably overlying Ghanzi Group represents an extensive rift enlargement episode.
The basal Kuke Formation contains siliciclastic sedimentary rocks interbedded with conglomerates
containing locally derived clasts of Kgwebe Formation volcanic rocks. This locally preserved sequence is
in turn overlain by ~2.5 kilometers of continental to shallow shelf siliciclastic sedimentary rocks
(Ngwako Pan Formation), ~1.5 kilometers of mixed-marine siliciclastic and carbonate rock (D’Kar
Formation), and ~1 to up to 4 kilometers of nearshore to fluvial siliciclastic and carbonate rocks
(Mamuno Formation; Litherland, 1982; Schwartz et al., 1995; Modie, 1996; Kampunzu et al., 2000;
Modie, 2000). Detrital zircon studies from the Ghanzi basin indicate that the combined middle to upper
Ghanzi Group was deposited after 1056 ± 9 Ma based on the youngest cluster of overlapping detrital
zircon ages (Chapter 3). Recent investigations into the tectonostratigraphic evolution of the basin have
revealed a previously unrecognized period of extension that occurred after deposition of the Ngwako Pan
Formation but prior to deposition of the D’Kar Formation. Extension resulted in paired uplift and
subsidence of fault blocks, with localized erosion and re-deposition of up to 100’s of meters of Ngwako
Pan Formation sedimentary rocks (Chapter 3).

In Namibia, the Mamuno Formation is tentatively correlated with the Kamtsas Formation, the
lower portion of the Nosib Group of the Damara Supergroup. Hegenberger (1993) described the Kamtsas
Formation in the Witvlei-Gobabis area of easternmost Namibia as being in unconformable angular
contact with the Doornpoort Formation (Ngwako Pan Formation equivalent). The Nosib Group is
overlain by the Witvlei Group, which consists of diamictites and cap carbonate rocks that are correlated
with the Cryogenian (~750 to ~600 Ma) ‘Snowball Earth’ events through chemostratigraphy
(Hegenberger, 1993; Prave et al., 2011). These data constrain the age of the Ghanzi Group to between
~1050 to ~750 Ma.

In Botswana, the Ghanzi Group is unconformably overlain by siliciclastic rocks of the Okwa
Group. These rocks were interpreted as a syn-orogenic molasse sequence deposited during the Pan-
African (~560 to ~ 480 Ma) Damara Orogen that folded the disconformably underlying Kgwebe
Formation and Ghanzi Group rocks; the resulting fold and thrust belt is referred to as the Ghanzi-Chobe
zone (Figure 4.1; Modie, 1996). The lower Kaçgae Subgroup, which is folded with the underlying basement rocks, yielded a detrital zircon with a U-Pb age of 579 ± 12 Ma (Ramokate et al., 2000). The U-Pb age constrains the timing of deposition of the Okwa Group as well as the timing of Pan-African deformation. Overlying strata of the upper Boitsevango Subgroup were gently titled and deposited in graben structures, indicating its deposition post-dates the Pan-African folding event (Ramokate et al., 2000).

The Ghanzi-Chobe zone separates Paleoproterozoic and Archean rocks of the Kalahari Craton from high-grade, penetrative deformed domains of the Damara Belt in northern Botswana (Figure 4.1; Schwartz et al., 1995; Modie, 1996; Miller, 2008). Recent interpretation of combined government geophysical surveys correlates this zone with the Southern Foreland zone of the Damara belt in Namibia (Figure 4.1; Miller, 2008; Lehmann et al., 2015). The belt is ~150 to 200-km-kilometers-wide in Botswana and is characterized by upright to inclined, doubly-plunging folds that have strike lengths of 35- to 60-km and wavelengths of ~5-12 kilometers; the fold axes define northwest to southeast shortening (Schwartz et al., 1995; Modie, 1996; Hall, 2013a; Lehmann et al., 2015). Although no dating has been undertaken in the Ghanzi-Chobe zone, the timing of the presumed folding event can probably be constrained by phyllites of the Southern Foreland Zone of the Damara orogen in Namibia that record apparent $^{40}$Ar/$^{39}$Ar mica crystallization ages of $\sim$568–553 Ma; however, these are maximum ages due to variable influence of detrital mica (Gray et al., 2006).

The volcano-sedimentary rock package of the Ghanzi-Chobe belt was metamorphosed to lower greenschist facies during regional deformation (Schwartz et al., 1995). K/Ar dating of muscovite and illite indicate three thermal events in the Southern Foreland Zone in Namibia with peak metamorphism at 530 Ma, a later event at 495 Ma, and final deformation at ~480 Ma (Ahrendt et al., 1978; Gray et al., 2006). In Botswana, major northeast-trending faults with right-lateral displacement cross-cut the folded Ghanzi Group rocks; one such fault was traced for 15-km along the northwestern flank of the Nguneakau Hills (Figure 4.1; Schwartz and Akanyang, 1994b). Both the regional folds and northeast-trending structures are in turn cut by series of regional-scale north-northeast-trending faults identified in aeromagnetic datasets; these structures also display right-lateral displacement in map view and have been noted to contain quartz veins (Schwartz and Akanyang, 1994b).

### 4.3.2 Sedimentary rock-hosted Cu-Ag deposits

Copper-silver deposits in the Kalahari Copperbelt occur within the lowermost D'Kar Formation. The deposits share many characteristics typical of sedimentary rock-hosted copper deposits, including locations along a major stratigraphic reduction-oxidation (redox) barrier that aided the precipitation of vertically- and laterally-zoned base metal sulfide minerals (chalcolite, bornite, chalcopyrite, pyrite, galena, sphalerite; Borg and Maiden, 1986a; 1896b, 1989; Borg, 1995; Schwartz et al., 1995; Modie,
Base metal sulfide minerals occur as disseminations as well as in cleavage-parallel lenticles, quartz-carbonate veins, and brittle to ductile shear fabrics (Borg and Maiden, 1986; Schwartz et al., 1995; Modie, 2000; Van der Heever et al., 2009; Hall, 2013a).

Quartz-carbonate veins host the majority of sulfide minerals. Hydrothermal quartz-carbonate veins within the D’Kar Formation are most prevalent within fine-grained, thin bedded to laminated siltstone-mudstone-marlstone. Veins range from mm-scale to several cm in width. Vein densities vary from one or less to greater than ten veins per meter. Most veins are bedding-parallel or sub-parallel and many are linked via discordant veinlets/veins/fractures (Davies, 2013; Hall, 2013a). Some discordant veins are oriented parallel to cleavage indicating the syn-kinematic nature of the vein systems. West-northwest- and north-northeast-trending discordant veins form part of conjugate vein systems that structural analysis suggests fit well with the overall southeast-directed compressional stress regime operation during Damara orogenesis (Hall, 2013a). A minor set of northwest-trending veins also occurs.

Reverse shear sense during and after the time of vein formation is indicated by slickenlines, S-C fabrics, and boudinage of bedding-parallel veins; discordant veins also display boudinage (Davies, 2013; Hall, 2013a). Cross-cutting relationships indicate that several generations of quartz-carbonate veins are present. Veins commonly show evidence for reactivation including crack-seal textures, brecciation and milling textures, and boudinage, which are cross-cut by undeformed veins (Hall, 2013a).

The bedding-parallel vein systems are most prevalent near the boundaries between layers of contrasting rheology (i.e. competent sandstone of the Ngwako Pan Formation adjacent to rheologically weak fine-grained siliciclastics and carbonates of the D’Kar Formation). The fractal nature of the rheological contrasts (bedding- to formation-scale) facilitated brittle-ductile deformation within the finer-grained lithologies (i.e. veins along bedding planes between sandstone beds and mudstone beds) of the lowermost D’Kar Formation. The under- and overlying rheologically competent lithologies (Ngwako Pan Formation and marine sandstone of the D’Kar Formation) underwent brittle failure during folding. These characteristics and abundant reverse-shear sense indicators (see above) suggest that the flexural-slip mechanism was operational during folding, with much of the strain, and consequently hydrothermal fluid flow, having been localized within the finer-grained units of the D’Kar Formation over prolonged periods of compressional folding (Davies, 2013; Hall, 2013a).

Alteration associated with the mineralizing hydrothermal fluids of the Kalahari Copperbelt is cryptic in most cases due to the fine-grained nature of the host rocks. Automated mineralogy analysis of samples from the Boseto District indicate that at least two generations of alteration are present within the host rocks (Hall, 2013). The first generation consists of a quartz-albite-chlorite-(sulfide) assemblage that forms thin, inconspicuous layers parallel to bedding. The second generation consisting of potassic alteration that is observed cross-cutting the earlier quartz-albite-chlorite-sulfide assemblage is intimately
associated with calcite-(dolomite-siderite-quartz-sulfide) veinlets and carbonate-cored shear structures (Hall, 2013a).

Fluid inclusion studies suggest that the mineralizing fluids were H$_2$O-NaCl ± CaCl$_2$ brines with salinities ranging between 4.0 and 25.33 wt. % NaCl equiv. with trapping temperatures between 150 and 350 °C (Schwartz et. Al., 1995; Hall, 2013a; Morgan, 2013; Shephard, 2014; Walsh, 2014). δ$^{34}$S values for sulfides throughout the Kalahari Copperbelt range from +5 to -55‰, with most values falling between -5 to -15‰ and -20 to -30‰, suggesting that the sulfur source for the mineralizing system(s) was primarily bacteriogenic in origin (Ruxton, 1981; Hall, 2013a; Morgan, 2013, Shephard, 2014; Walsh, 2014).

4.4 Sample descriptions and petrography

Mineralized drill core containing vein-hosted and disseminated molybdenite and molybdenite-bornite, vein-hosted chalcocite-idaite, and cleavage-hosted arsenopyrite from selected deposits was logged, and samples were taken for petrographic and isotopic analysis (Table 4.1, see Figures 4.1 for map locations and Figure 4.2 for drill hole locations). Polished thin sections of the samples were examined in both transmitted and reflected light to characterize mineral assemblages and textures. The samples were further analyzed on a TESCAN MIRA3 LMH Schottky field emission-scanning electron microscope (operated at 15 and 25 kV) equipped with a Bruker XFlash® 6/30 silicon drift detector for energy-dispersive X-ray spectrometry (EDS) at Colorado School of Mines.

4.4.1 Northeast Fold deposit, molybdenite (AIRIE Run # MD-1700)

Molybdenite bearing veins occur within a massive sandstone bed of the Ngwako Pan Formation within the crest of the Banana Zone anticline at the Northeast Fold deposit (Figure 4.2). The sandstone is cut by structurally-controlled alteration that ‘bleached’ the pink-reddish host rock sandstone. Bleached zones are in turn cross-cut by several generations of veins including quartz-calcite-(sulfide), quartz-calcite-potassium feldspar-(sulfide), and calcite-only veins. Semi-massive replacement of wall rock adjacent to veins by molybdenite intergrown with bornite-(digenite) occurs along both concordant and discordant fractures that are spatially and temporally related to quartz-calcite-potassium feldspar veins (Figure 4.3a).

Concordant fractures are rimmed by coarse-grained molybdenite intergrown with lesser bornite (Figure 4.3, b-d) that contain irregular remnants of sandstone. Back-scatter electron (BSE) imagery indicates that the molybdenite-(bornite) mixture is relatively massive adjacent to fractures. In addition to molybdenite and bornite, minor wittichenite (Cu$_3$Bi$_3$S$_4$), potassium feldspar, and monazite form irregular patches along the boundaries between molybdenite and bornite (Figure 4.3c). Silver was detected in EDS analysis as a minor constituent within wittichenite. Some euhedral grains of molybdenite contain elevated Re concentrations detectable by EDS analysis (Figure 4.3d). Minor molybdenum was also
<table>
<thead>
<tr>
<th>District</th>
<th>Deposit/ Prospect</th>
<th>Borehole ID</th>
<th>Depth (m)</th>
<th>Formation</th>
<th>Host rock</th>
<th>Sample type</th>
<th>Sulfide minerals present*</th>
<th>Gangue minerals*</th>
<th>AIRIE Run # (Age)</th>
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<tr>
<td>Banana Zone</td>
<td>Northeast Fold</td>
<td>HA-483-D</td>
<td>158.4</td>
<td>Ngwako Pan</td>
<td>Coarse-grained lithic-arkose, weakly bleached</td>
<td>Fracture-to vein-hosted Mo-Bn</td>
<td>Mo, Bn, Cc-Dg, Wit</td>
<td>Qtz, Cal, Kfs, Rt, Mnz</td>
<td>MD-1700</td>
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<tr>
<td></td>
<td>Northeast Mango Two</td>
<td>MGDD2408</td>
<td>190.0</td>
<td>D’Kar</td>
<td>Thin bedded to laminated marlstone-siltstone-mudstone</td>
<td>Veinlet- to disseminated Mo</td>
<td>Mo, Cp, Gt, Sp, Sph, Wit, Bism, Ni-Co-Fe sulfarsenide</td>
<td>Qtz, Cal, Dol, Ank, Chl, Rt, *Xnt, Aln</td>
<td>MD-1370, MD-1380</td>
</tr>
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<td>501.1</td>
<td>D’Kar</td>
<td>Laminated, non-calcareous siltstone</td>
<td>Vein-hosted Cc-Dg-Bn</td>
<td>Ce, Id, Bn, Dg, Cp, Ga, Sp, (Cd-, Hg-, and Se-rich inclusions)</td>
<td>Qtz, Cal, Chl, Bt, Rt, Kfs, *Xnt, Aln, REE-(fluoro-) carbonates</td>
<td>LL-957</td>
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<td>Northeast Zone 5</td>
<td>HA-1049-D</td>
<td>1085</td>
<td>D’Kar</td>
<td>Interlaminated siltstone to calcareous siltstone</td>
<td>Cleavage lenticle-hosted Asp</td>
<td>Asp, Cp, Sp, Gn, Ni-Co-Fe sulfarsenide</td>
<td>Qtz, Cal, Chl, Rt, REE-(fluoro-) carbonates</td>
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<td>68.6</td>
<td>Kgwebe</td>
<td>Rhyolitic volcanoclastic</td>
<td>Vein-hosted Cp</td>
<td>Cp</td>
<td>Mgt, Qtz, Cal, Ab, Kfs, Bt, Ep, Chl Tnt, Br, Ap, REE-silicates</td>
<td>MD-1362**</td>
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<tr>
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<td>Vein-hosted Bn</td>
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<td>Qtz, Cal, Chl, Bt, Ms/Phg,</td>
<td>LL-610′ (1012 ± 17 Ma)</td>
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<tr>
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<td>Vein-hosted Cp</td>
<td>Cp, Mo</td>
<td>Qtz, Cal, Chl, Bt, Ms/Phg, Hem, Rt, Xnt, REE-(fluoro-) carbonates</td>
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<tr>
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Abbreviations (tables and text): Ab = albite; Ag = native silver; Aln = allanite; Ank = ankerite; Apt = apatite; Asp = arsenopyrite; Bis = bismuthinite; Bn = bornite; Bt = barite; Bi = biotite; Cal = calcite; Cc = chalcocite; Chl = chlorite; Cp = chalcopyrite; Dg = digenite; Dol = dolomite; Ep = epidote; Ga = galena; Hem = hematite; Kfs = potassium feldspar; Mgt = magnetite; Mnz = monazite; Mo = molybdenite; Ms = muscovite; Pg = phenigite; Qtz = quartz; REE = rare earth elements; Rt = rutile; Sp = sphalerite; Tnt = titanite; Wit = wittichenite; Xnt = xenotime

* determined from reflected light and scanning electron microscopy with electron dispersive spectrometry

** sample did not yield measurable amounts of Re for isotopic analysis, see Appendix A-1 for more detailed information and petrography

† Re-Os age reported in Hall (2013), sample LL-610 was marginally LLHR with minor common Os; the error reflects the use of different initial Os compositions (a range of ages) rather than true error for a precise age. See Appendix A-3 for additional information and petrography on the LLHR sample LL-611. Sample LL-612 contained high levels of common Os (not LLHR) resulting in a wider age range dependent on reasonable initial Os composition.
Figure 4.2 Generalized cross-sections of Kalahari Copperbelt Cu-Ag deposits highlighting ore zones and the locations of samples utilized in Re-Os and U-Th-Pb isotopic analysis.
Figure 4.3 HA-483-D 186.6 to 193.2 meters, fracture- to vein-hosted molybdenite-bornite.  A) Molybdenite (large arrow) used for Re-Os geochronology passes into bornite-digenite-chalcocite along the discordant linking fracture plane and connects with a bedding-parallel (concordant) quartz-calcite-potassium feldspar-bornite-molybdenite vein.  B) Reflected light image of intergrown molybdenite (silver-white color), rutile (bright white), and bornite with thin digenite rims.  C) BSE image showing coarse-grained molybdenite intergrown with irregular bornite (darker gray), wittichenite (white) and potassium feldspar (dark).  The wittichenite may contain minor amounts of Ag.  D) BSE image displaying irregular to crystalline molybdenite intergrown with bornite (darker gray) with chalcocite rims, potassium feldspar (black) and a tabular potassium-sodium-magnesium aluminosilicate phase.  The crystalline molybdenite contains a Re-rich inclusion.  E) BSE image showing disseminated molybdenite grains intergrown with quartz, potassium feldspar, and rutile.
detected within bornite along molybdenite-bornite grain boundaries, implying rapid precipitation of the coeval sulfide phases. Grains of molybdenite (Figure 4.3e) that occur away from the fracture may contain minor inclusions of monazite, bornite-chalcocite, wittichenite, and an unidentified Cd-Cu-S mineral phase. Potassium feldspar and minor rutile are ubiquitous intergrowths with molybdenite and bornite (Figure 4.3e). Minor amounts of a tabular shaped (Na-Mg)-bearing potassium aluminosilicate phase are also intergrown with the molybdenite-bornite assemblage (Figure 4.3d) and may reflect incomplete potassic alteration of Na-(Mg)-bearing aluminosilicate minerals.

4.4.2 Northeast Mango Two deposit, molybdenite (AIRIE Run #’s MD-1370 and MD-1380)

A high Mo assay value sample consisting of thinly bedded and normally graded calcareous siltstone with a thin interbed of dark gray-blue mudstone from the Northeast Mango Two deposit was selected for Re-Os geochronology (Figure 4.4a). The sample contains a bedding-parallel quartz-calcite-chalcopyrite vein along the contact between siltstone and mudstone. Mineralized clasts and a deformed seam of mineralized mudstone occur within the vein. The quartz-calcite-chalcopyrite vein is cut by a shear fabric, which is in turn cut by discordant, vuggy textured veinlets (Figure 4.4a).

Petrographic analysis reveals that the mudstone bed is composed of carbonate-, siliciclastic-, and phengitic mica-rich ripple laminations that display reactivation surfaces (Figure 4b). Finely disseminated sulfide grains are ubiquitous throughout the mudstone bed but are concentrated within siliciclastic laminations (Figure 4.4b-c). Molybdenite is the primary disseminated sulfide mineral. It occurs as 5-50µm wide, ovoid- to irregularly-shaped, matted to felt-like textured aggregates that are texturally similar to framboidal pyrite (Figure 4d) as well as smaller intergranular patches and individual micron-sized bladed grains. The molybdenite aggregates commonly contain micron-sized inclusions of galena, wittichenite (CuBiS₃), bismuthinite (BiS₂), and Ni-Co-Fe sulfarsenide minerals (Figure 4.4d); the latter also forms subhedral crystals intergrown with the molybdenite aggregates. The phengitic mica-rich laminations contain 2-5 volume % disseminated molybdenite as 5-10 µm aggregates and individual grains, as well as distinctive ring- or atoll-shaped rims to rounded quartz and carbonate grains (Figure 4.4e); the texture is suggestive of sulfide replacement of organic matter-rich rims or grain coatings.

The host rock is cut by a 4.5-cm-thick quartz-calcite-chalcopyrite vein that contains fragments of the mudstone bed (Figure 4.4). These fragments are encased by blocky, often euhedral and terminated, quartz crystals that indicate growth into open space. The vein quartz is fractured and infilled by coarse crystalline calcite. Coarse blebs of chalcopyrite are intergrown with and commonly occur at the interface between quartz and calcite. Minor galena, sphalerite, wittichenite, bismuthinite, molybdenite, and (Ni-Co-Fe) sulfarsenide occur at the margins of and are intergrown with the chalcopyrite (Figure 4.5a-b). Xenotime (YPO₄) and allanite ((Ce, Ca, Nd)₂(Al, Fe³⁺)₃(SiO₄)₃(OH)) occur as inclusions in the chalcopyrite (Figure 4.5).
Figure 4.4 Sample MGDD2408_190 m, veinlet to disseminated molybdenite. A) Macro-photo of the quartz-calcite-chalcopyrite vein with molybdenite-bearing mudstone on the lower edge and a clast of molybdenite-bearing mudstone encased within the vein. The upper margin of the vein displays a strong shear texture infilled with chalcopyrite. The vein is cut by a late calcite-pyrite veinlet. B) BSE image of the undeformed mudstone bed with ripple laminations composed of phengite, quartz-chlorite, and carbonate minerals. The abundant bright white spots are disseminated sulfides, primarily molybdenite. C) Reflected light image showing abundant disseminated molybdenite (gray) and coarser chalcopyrite clots along a bedding plane. D) BSE image displaying a framboid-like molybdenite cluster with abundant inclusions of galena and bismuthinite. E) BSE image displaying finely disseminated (micron-size) molybdenite laths and ring- to atoll-like molybdenite rims on quartz-carbonate grains in a phengitic layer; the textures are suggestive of replacement of organic material.
Figure 4.5 Sample MGDD2408_190 m, veinlet to disseminated molybdenite. A) Reflected light image displaying a hairline Mg-calcite-galena veinlet with a molybdenite selvage. Disseminated molybdenite occurs outboard of the veinlet along with chalcopyrite. B) BSE image displaying a dolomite-calcite-molybdenite-bismuthinite vein with a selvage of molybdenite. C) Reflected light image displaying galena, wittichenite, and bismuthinite within calcite adjacent to the chalcopyrite. D) BSE image displaying molybdenite intergrown with subhedral (Ni, Co, Fe) sulfarsenide and chalcopyrite.

Both the intact mudstone bed and the vein-hosted mudstone fragments contain smaller 10-µm-wide veinlets composed of Mg-calcite, zoned Fe-and Mn-dolomite, minor quartz and rutile. These veinlets also contain molybdenite, galena, chalcopyrite, pyrite, wittichenite, and bismuthinite and may have thick selvages composed of massive molybdenite (Figure 4.5c-d). Cross-cutting relationships between these micro-veinlets and the large quartz-calcite-chalcopyrite vein could not be established within the sample.

The upper margin of the quartz-calcite vein is cut by a shear fabric formed by mica and displays boudin-like textures in quartz and reverse-sense shear fabrics containing abundant calcite and chalcopyrite (Figure 4.4a). Straight to bent deformation twins in calcite also indicate post-crystallization deformation. Both the quartz-calcite vein and the micaeous shear fabric are cut by a discordant, vuggy textured veinlet that contains fragments of molybdenite mineralized wall rock and quartz-calcite-
chalcopyrite vein. EDS analysis indicates the core of this veinlet is composed of Mg-calcite with minor rare earth-bearing apatite and traces of silver while the margins are composed of Ca-Zr-(Ti) silicate mineral phases (gittensite-baghdadite) intergrown with the Mg-calcite.

4.4.3 Zone 6 prospect, chalcocite-idaite (AIRIE Run # LL-957)

A 2-cm-thick bedding-parallel quartz-calcite-chalcocite-idaite-bornite-digenite-native silver vein cutting laminated, non-calcareous siltstone was selected for Re-Os geochronology (Figure 4.6a). The vein is composed of fractured blocky quartz that is infilled by calcite and sulfide minerals; angular to partially resorbed fragments of quartz encased by sulfide and calcite are present within the core of the vein (Figure 4.6a-b). The lower portion of the vein consists primarily of intergrown chalcocite-idaite-digenite (Figure 4.6c) while the upper margin of the vein contains intergrown bornite-idaite-digenite with xenotime and minor barite present along the upper margin of the vein. Rutile occurs as small, discrete crystals intergrown with calcite and quartz. A small discordant veinlet present below the vein contains abundant barite and Mn-calcite with minor chalcopyrite, rutile, and a Sr-Y-Ca-Na carbonate mineral that EDS elemental abundances suggest is donnayite-(Y). Potassium feldspar is present in the discordant veinlet and replaced part of the adjacent wall rock (Figure 4.6).

The vein is surrounded by disseminated chalcocite that gives way to a bornite-chalcopyrite assemblage further from the vein margins. Pb, Zn, Ag, Cd, Hg, and/or Se are present in micron-sized inclusions within the sulfides. Rutile is a commonly intergrown with the disseminated sulfide minerals. LREE-bearing mineral phases intergrown with larger disseminated grains of chalcocite-bornite include blebs allanite-(Ce), acicular parasite-(Ce), and donnayite-(Y) (Figure 4.6d-e). Textures suggests that albite was replaced by copper sulfide minerals, potassium feldspar, and LREE silicates and fluoro-carbonate minerals adjacent to the vein (Figure 4.6e).

4.4.4 Zone 5 deposit, arsenopyrite (AIRIE Run #’s LL-934 and LL-962)

A deformed interlaminated siltstone and calcareous siltstone containing euhedral, 0.5 – 4.0 mm long arsenopyrite crystals was sampled for Re-Os geochronology (Figure 4.7a). The host rock contains disseminated sphalerite and is cut by bedding-parallel to wispy quartz-calcite-sulfide veinlets containing chalcopyrite, sphalerite as well as minor galena and hematite. The host rock displays a spaced cleavage ($S_1$; Figure 4.7a) developed at an acute angle to bedding. Cleavage-parallel quartz-calcite-(sulfide) lenticles are developed between bedding-parallel veins. A small-scale fault propagation fold with its axial plane and micro-fault oriented parallel to $S_1$ occurs within one lamination; the core of micro-fault is infilled by an assemblage of quartz-calcite-sphalerite-galena-chalcopyrite (Figure 4.7a).

Arsenopyrite crystals in non- or weakly sheared domains have symmetric quartz-filled pressure fringes (red arrows indicate directions of fibrous quartz growth in Figure 4.7a). Within areas with well-developed S-C fabrics, these pressure fringes are highly rotated (orange arrows in Figure 4.7b) and new,
asymmetric pressure fringes were filled by quartz (red arrows in Figure 4.7b). This indicates that the arsenopyrite crystals and the pressure fringes were affected by post-crystallization deformation related to cleavage and S-C fabric development (Figure 4.7c). Within the S-C fabric, some pressure shadows contain an outer rim of quartz infilled by calcite and minor galena, sphalerite, and chalcopyrite that have nucleated on the crystal faces of the arsenopyrite (Figure 4.7d). These sulfides are commonly associated with and/or intergrown with minor amounts of a bladed to acicular habit mineral; elemental abundances from EDS analysis indicate it is a REE-bearing fluoro-carbonate mineral, probably parasite-(Nd) \((\text{Ca}(\text{Ce},\text{La},\text{Nd})_2(\text{CO}_3)_3\text{F}_2)\). Most of the larger arsenopyrite crystals display minor boudinage with calcite-filled fractures oriented near perpendicular to \(S_1\) (Figure 4.7e). Sphalerite and Ni-Co-Fe sulfarsenide replace arsenopyrite along and adjacent to these transtensional fractures (Figure 4.7f-g).
Figure 4.7 Sample HA-1049-D_1085 m, siltstone with veinlets and coarse arsenopyrite crystals. A) Hand specimen image showing laminated siltstone with euhedral arsenopyrite crystals and quartz-calcite-(sphalerite-galena-chalcopyrite) cleavage lenticles. Note the presence of an $S_1$ (yellow) cleavage resulting in jagged bedding planes. B) Polarized light image of a coarse-grained, prismatic arsenopyrite crystal with well-developed feathery quartz pressure fringes (red arrows) displaying growth perpendicular to crystal faces. C) Combined reflected and polarized light image of a euhedral arsenopyrite crystal with highly rotated and recrystallized fibrous quartz pressure shadows. The rotation (orange arrows) is related to a S-C fabric. Red arrows point to secondary quartz pressure shadows developed adjacent to the arsenopyrite. D) Reflected light image displaying a sphalerite-chalcopyrite-galena assemblage intergrown with calcite adjacent to a prismatic arsenopyrite crystal. Note how the galena overgrows the arsenopyrite in the bottom left. E) Reflected light image of tensional fractures (boudinage, blue arrows) of the arsenopyrite crystal in (C). F) Galena and Ni-Co-Fe sulfarsenide replacing arsenopyrite along micro-fractures. G) BSE image displaying overgrowth and/or replacement of coarse-grained arsenopyrite by sphalerite adjacent to one of the extensional fractures perpendicular to $S_1$. EDS analysis indicates that the sphalerite adjacent to the arsenopyrite contains up to one modal weight percent As. The sphalerite is associated with ferroan and manganoan calcite.

4.5 Methods

Rhenium-osmium geochronology, sulfur stable isotopic analyses, and U-Th-Pb xenotime geochronology were carried out on the samples from the Kalahari Copperbelt. The methodology for each of these techniques is described below.
4.5.1 Re-Os geochronology

Re-Os geochronology was carried out by the A.I.R.I.E. Program at Colorado State University. Sulfide separates were obtained using a hand drill and/or through hand-crushing and picking of grains under a binocular microscope. Sulfide separates were weighed and then combined with $^{185}$Re and $^{190}$Os spikes. Due to high Re concentrations, molybdenite was combined with a mixed-double Os spike (denoted by the prefix MD-), whereas other mineral separates only required a single Os spike (denoted by the prefix LL-). All samples and spike were digested and equilibrated in HNO$_3$-HCl (inverse aqua regia) using the Carius tube method (Shirey and Walker, 1995). Once Re and Os were chemically isolated and purified, the clean fractions of each were separately loaded onto Pt filaments and Re and Os isotopic ratios were measured by negative thermal ionization mass spectrometry (NTIMS) on a Triton machine. Isotopic analysis utilizing NTIMS dramatically increases the ionization of Re and Os, enabling precise measurements of Re and Os isotopic compositions on pictogram quantities (Stein, 2014). When employed on molybdenite or low level, high radiogenic ( LLHR) sulfide species (Stein et al., 1997, 2001; Stein et al., 2000), the method yields highly accurate and precise ages (e.g. Chesley and Ruiz, 1998). Age calculations used the decay constant for $^{187}$Re determined by Smoliar et al. (1996).

4.5.2 Sulfur stable isotopic analysis

Initial sulfur isotopic analysis of sulfide minerals was directed at determining the source of sulfur for both the Re-Os samples. Sulfide separates were obtained using a dental drill. Approximately 20-25 μg of an individual sample was combusted in a Eurovector 3000 elemental analyzer, yielding sulfur dioxide that was delivered to the mass spectrometer using continuous-flow techniques with helium as the carrier gas. Values of $\delta^{34}$S are expressed relative to the Vienna Cañon Diablo Troilite (VCDT) standard, using the NBS-127 standard reference material obtained from the National Institute of Standards and Technology. Repeat analysis of a lab-working standard (also barium sulfate) yielded a precision of 0.5‰. The mean half range for duplicates is 0.25‰.

4.5.3 Xenotime U-Th-Pb LASS ICP-MS geochronology

Xenotime grains were identified in two of the samples (MGDD2408_190; HA-664-D_501.1) that were dated using Re-Os geochronology. A single xenotime grain in sample MGDD2408_190 (AIRIE run #’s MD-1370 and MD-1380; Figure 4.8a-b) was identified with Energy Dispersive X-Ray Spectroscopy (EDS) based on the yttrium and phosphorus peaks in the EDS spectrum. It occurs as a small, irregularly shaped inclusion within a large chalcopyrite grain that also contains inclusions of molybdenite, sphalerite, bismuthinite, emplectite, galena, and arsenopyrite. BSE imaging shows that the xenotime inclusion contains faint compositional zoning (Figure 8b).
Multiple xenotime grains in sample HA-664-D_501.1 (AIRIE run # LL-957) were identified with EDS. They occur as small (5-20 µm), rounded to blocky to irregularly shaped grains within the wall rock immediately adjacent to the upper margin of the quartz-calcite-chalcosite-idaite-bornite-digenite-native silver vein (Figure 4.8c). The grains are commonly overgrown and partially replaced by bornite-digenite (Figure 4.8d). BSE imaging of the xenotime grain clusters did not reveal any distinct internal zonation within the individual grains.

Prior to LA-ICPMS analysis, the trace-element geochemistry of the xenotime grain in sample HA-664-D_501.1 was determined utilizing electron probe microanalysis through wavelength dispersive spectrometry (WDS). The analyses were carried out on a JEOL 8900 Electron Microprobe housed at the
USGS Denver Microbeam Laboratory in Denver, Colorado, operated at 20 kV with a Faraday cup beam current of 50 nA. The electron-probe data was gathered to compare to the LA-ICPMS data to ensure the same spots were being analyzed by each method.

LA-ICPMS spot analyses of xenotime were conducted at the University of California Santa Barbara laser-ablation split-stream petrochronology laboratory, using a Photon Machines 193 nm laser, coupled to a Nu Instruments Plasma HR (for U-Th-Pb geochronology) and an Agilent 7700 quadrupole (element concentrations), following the methods of Kylander-Clark et al. (2013). Elemental abundances were used solely for ensuring data quality; count rates of $^{31}$P, coupled with Y and REE calculated abundances ensured that the xenotime targeted by electron-beam methods was the same that was analyzed via LA-ICPMS. Continuous collection of laser-ablation data occurred over a period of 20 seconds at 4 Hz and ~1 J/cm$^2$ for each individual spot analysis. The FC1 xenotime (56.5 Ma) was interspersed throughout the run and used as the primary reference material to reduce U-Th-Pb data. Elemental concentrations were calculated using Stern monazite, a relatively homogenous, in-house reference material; P was used as the internal standard. LA-ICPMS data was processed with Iolite v 2.5 (Paton et al., 2011), and plotted on Tera-Wasserburg concordia diagrams using Isoplot 4.1 (Ludwig, 2008).

4.6 Results

The results of the Re-Os geochronology, sulfur isotopic analysis of the Re-Os sulfide samples, and U-Th-Pb xenotime geochronology are presented below. Additional petrography and inconclusive Re-Os results are included in Appendix A.

4.6.1 Re-Os geochronology

A 15.91 mg molybdenite-bornite separate from the Northeast Fold deposit (Figure 4.3; AIRIE run # MD-1700) contained an extremely high Re concentration of nearly 2000 ppm. The sample was combined with mixed double Os spike ($^{185}$Re and $^{190}$Os) and returned a direct, single-sample Re-Os age of 515.9 ± 2.4 Ma (Table 2).

Two molybdenite separates with sample weights of 21.3 mg (AIRIE run # MD-1370) and 21.8 mg (AIRIE run # MD-1380) and Mo concentrations of 60% and 70%, respectively, were obtained from the vein-hosted mudstone seam of sample MGDD2408_190. The molybdenite separate may have contained trace molybdenite from the adjacent wall rock as well as the vein selvage. Sample MD-1370 was underspiked due to the high concentration of Re in the sample (256 ppm, Table 4.2). Molybdenite from both separates yielded indistinguishable single sample Re-Os ages of 981 ± 8 (MD-1370) and 981 ± 3 Ma (MD-1380; Table 4.2). The larger uncertainty in sample MD-1370 is due to the sample being underspiked. The 2σ uncertainties in the calculated ages primarily reflect the uncertainty in the $^{187}$Re decay constant (0.31%) and to a much lesser extent the uncertainty in the $^{185}$Re and $^{190}$Os spike calibrations (0.05% and 0.06%, respectively).
Table 4.2 Re-Os dating, mixed-double Os spike molybdenite, Kalahari Copperbelt

<table>
<thead>
<tr>
<th>AIRIE Run #</th>
<th>Sample ID (Drill Hole ID, depth, prospect, sulfide)</th>
<th>Re, ppm</th>
<th>2σ</th>
<th>187Os, ppb</th>
<th>2σ</th>
<th>Common Os, ppb</th>
<th>2σ</th>
<th>Age, Ma</th>
<th>2σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD-1700</td>
<td>HA-483-D, 158.4 m, “Northeast Fold” molybdenite</td>
<td>1825.8</td>
<td>5.8</td>
<td>9905.7</td>
<td>7.6</td>
<td>0.032</td>
<td>0.049</td>
<td>515.9</td>
<td>2.4</td>
</tr>
<tr>
<td>MD-1370</td>
<td>MGDD-2408, 190.0 m, “Northeast Mango Two” molybdenite</td>
<td>256</td>
<td>2</td>
<td>2654</td>
<td>2</td>
<td>0.085</td>
<td>0.004</td>
<td>981.0</td>
<td>8</td>
</tr>
<tr>
<td>MD-1380</td>
<td>MGDD-2408, 190.0 m, &quot;Northeast Mango Two&quot; molybdenite (b)</td>
<td>185.3</td>
<td>0.2</td>
<td>1918</td>
<td>2</td>
<td>0.06</td>
<td>0.02</td>
<td>981.0</td>
<td>3</td>
</tr>
</tbody>
</table>

All uncertainties reported at two-sigma; uncertainty in Re-Os ages includes the 187Re decay constant uncertainty
Re-Os data acquired by Carius tube dissolution and sample equilibration with a double Os spike for mass fractionation and determination of common Os
Molybdenite sample weights are 1.591 mg for MD-1700, and 21.3 and 21.8 mg for MD-1370 and MD-1380, respectively

Re blank for MD-1700 = 5.510 ± 0.01 pg, Os blank = 0.791 ± 0.008 pg with 187Os/188Os = 0.420 ± 0.004
MD-1370 is underspiked, and spiking was corrected for MD-1380, nevertheless, ages for these two runs are in excellent agreement; (b) designates a new mineral separate for a second analysis
Silicate dilution does not affect Re-Os ages

Table 4.3 Re-Os age determinations, single spike arsenopyrite and Cc-Id-Bn-native Ag mixture, D’Kar Formation, Kalahari Copperbelt

<table>
<thead>
<tr>
<th>AIRIE run #</th>
<th>Drill hole, deposit, sulfide sampled</th>
<th>Re, ppb</th>
<th>2σ</th>
<th>Total Os (ppb)</th>
<th>2σ</th>
<th>187Re/188Os</th>
<th>2σ</th>
<th>187Os/188Os</th>
<th>2σ</th>
<th>Assumed 187Os/188Os initial</th>
<th>Assigned Error</th>
<th>Model Age (Ma)</th>
<th>Error with λ187Re uncertainty (Ma)</th>
<th>Common Os (ppb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LL-957</td>
<td>HA-664-D, Zone 6 Cc-Id</td>
<td>27.088</td>
<td>0.112</td>
<td>0.1581</td>
<td>0.0028</td>
<td>79,093.8</td>
<td>1260.3</td>
<td>726.91</td>
<td>12.013</td>
<td>0.200</td>
<td>0.01</td>
<td>549.0</td>
<td>11.2</td>
<td>0.0017</td>
</tr>
<tr>
<td>LL-934</td>
<td>HA-1049-D, Zone 5 Asp</td>
<td>5.702</td>
<td>0.010</td>
<td>0.0616</td>
<td>0.0004</td>
<td>1321.8</td>
<td>7.4</td>
<td>15.168</td>
<td>0.109</td>
<td>0.800</td>
<td>0.01</td>
<td>648.9</td>
<td>8.6</td>
<td>0.0244</td>
</tr>
<tr>
<td>LL-962</td>
<td>HA-1049-D, Zone 5 Asp (b)</td>
<td>7.305</td>
<td>0.013</td>
<td>0.0764</td>
<td>0.0005</td>
<td>1471.7</td>
<td>8.2</td>
<td>16.936</td>
<td>0.121</td>
<td>0.800</td>
<td>0.01</td>
<td>654.5</td>
<td>8.5</td>
<td>0.0280</td>
</tr>
</tbody>
</table>

Arsenopyrite crystals of single geologic occurrence hand-picked from gently crushed fine-grained, markedly foliated, metasedimentary rock; (b) designates a new mineral separate for a second analysis
Bn = bornite, Cal = calcite; Cc = chalcocite, Id = idaite, Qtz = quartz
All uncertainties reported at two-sigma; uncertainty in Re-Os ages includes the 187Re decay constant uncertainty
Re-Os isotopic data acquired using single Re and Os spikes and a Carius tube aqua regia dissolution; isotopic ratios measured by NTIMS, AIRIE Program, Colorado State University
Assumed initial 187Os/188Os for age calculations for LL-934 and LL-962 = 0.8; assumed initial 187Os/188Os for age calculation for LL-957 = 0.2; assumed Os initial includes 10% uncertainty
For LL-934, Re blank = 0.996 ± 0.374 pg, Os blank = 0.077 ± 0.004 pg with 187Os/188Os = 0.326 ± 0.019
For LL-962 and LL-957, Re blank = 1.05 ± 0.078 pg, Os blank = 0.338 ± 0.010 pg with 187Os/188Os = 0.275 ± 0.011
Blank contribution from Re = 0.1%; blank contribution from Os = 0.4%; blank is insignificant to the ages calculated in this table
Table 4.4 Effect on calculated model age utilizing different initial Os compositions, single spike arsenopyrite samples, Zone 5 deposit

<table>
<thead>
<tr>
<th>AIRIE Run #</th>
<th>Assumed (^{187}\text{Os}/^{188}\text{Os}) initial ratio</th>
<th>Assumed error on Os initial ratio (10%)</th>
<th>Calculated Model Age (Ma)</th>
<th>Uncertainty on Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LL-934</td>
<td>0.200</td>
<td>0.02</td>
<td>675.9</td>
<td>8.2</td>
</tr>
<tr>
<td>LL-962</td>
<td>0.200</td>
<td>0.02</td>
<td>678.7</td>
<td>8.2</td>
</tr>
<tr>
<td>LL-934</td>
<td>0.400</td>
<td>0.04</td>
<td>666.9</td>
<td>8.2</td>
</tr>
<tr>
<td>LL-962</td>
<td>0.400</td>
<td>0.04</td>
<td>670.6</td>
<td>8.2</td>
</tr>
<tr>
<td>LL-934</td>
<td>0.600</td>
<td>0.06</td>
<td>657.9</td>
<td>8.4</td>
</tr>
<tr>
<td>LL-962</td>
<td>0.600</td>
<td>0.06</td>
<td>662.6</td>
<td>8.3</td>
</tr>
<tr>
<td>LL-934</td>
<td>0.800</td>
<td>0.08</td>
<td>648.9</td>
<td>8.6</td>
</tr>
<tr>
<td>LL-962</td>
<td>0.800</td>
<td>0.08</td>
<td>654.5</td>
<td>8.5</td>
</tr>
<tr>
<td>LL-934</td>
<td>1.000</td>
<td>0.10</td>
<td>640.0</td>
<td>8.9</td>
</tr>
<tr>
<td>LL-962</td>
<td>1.000</td>
<td>0.10</td>
<td>646.4</td>
<td>8.8</td>
</tr>
<tr>
<td>LL-934</td>
<td>1.200</td>
<td>0.12</td>
<td>631.0</td>
<td>9.3</td>
</tr>
<tr>
<td>LL-962</td>
<td>1.200</td>
<td>0.12</td>
<td>638.4</td>
<td>9.1</td>
</tr>
</tbody>
</table>

Re-Os ages calculated using six different assumed \(^{187}\text{Os}/^{188}\text{Os}\) initial ratios (ranging from 0.2 to 1.2) are shown in Table 4.4. Based on other Neoproterozoic mudstones (e.g. AIRIE unpublished, Kendall et al., 2006, Rooney et al., 2011, 2013, 2015), the \(^{187}\text{Os}/^{188}\text{Os}\) initial is most likely between 0.54 and 1.0 (corresponding to an age between 670-640 Ma).

A 49.651 mg chalcocite-idaite mixture from the Zone 6 quartz-calcite-chalcocite-digenite-bornite-native Ag vein was used for Re-Os geochronology (AIRIE run # LL-957). The single spike method returned very high \(^{187}\text{Re}/^{188}\text{Os}\) ratios with essentially no common Os, making it an LLHR composition. The analysis yielded a single sample mineral age of 549 ± 11 Ma (Table 4.3).

A 313-milligram sample of a coarse-grained arsenopyrite grain from the Zone 5 deposit (AIRIE run # LL-934) contained Re and Os levels that indicate a marginal LLHR character with the osmium radiogenic/common ratio about 2:1 (Table 4.3), making blank corrections insignificant. The sample was spiked and yielded a single sample mineral age of 648.9 ± 8.9 Ma. A second mineral separate (AIRIE run # LL-962), which contained a small amount of common osmium, returned a similar single sample age of 645.5 ± 8.5 Ma (Table 4.3). Although the sample requires further analyses to define the initial Os composition and a precise age through the isochron method (minimum of five analyses), a range of ages can be obtained by varying the assumed initial Os composition. The effect of utilizing different initial Os ratios of 0.2 to 1.2 for the model age (i.e. a mantle-like versus weathered continental crust source for Os) results in the calculated ages for samples LL-934 and LL-962 varying from approximately 630 Ma (initial Os ratio of 1.2) to 670 Ma (initial Os ratio of 0.2; Table 4.4).

4.6.2 Sulfur stable isotopic analysis

Sulfur stable isotopic results for all samples utilized for Re-Os geochronology were negative, ranging from −8.2 to −38.3‰ (Table 4.5). The molybdenite-bornite mixture from sample HA-483-D_158.4 yielded a \(\delta^{34}\text{S}\) value of −11.6‰. A \(\delta^{34}\text{S}\) value of −8.2‰ was obtained for the molybdenite in sample MGDD2408_190. The Zone 6 vein-hosted chalcocite-idaite mixture yielded a strongly negative \(\delta^{34}\text{S}\) value of −38.3‰. A \(\delta^{34}\text{S}\) value of -18.3‰ was obtained for the coarse-grained arsenopyrite grains from Zone 5.
Table 4.5 Sulfur stable isotopic analyses results

<table>
<thead>
<tr>
<th>Sample (drill hole, depth)</th>
<th>Deposit</th>
<th>Sulfide</th>
<th>Re/Os model age (Ma)</th>
<th>δ³⁴S (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HA-483-D, 158.4 m</td>
<td>Northeast Fold</td>
<td>molybdenite</td>
<td>515.9 ± 2.4</td>
<td>-11.6</td>
</tr>
<tr>
<td>MGDD-2408, 190.0 m</td>
<td>Northeast Mango Two</td>
<td>molybdenite</td>
<td>981.0 ± 3</td>
<td>-8.2*</td>
</tr>
<tr>
<td>HA-664-D, 501.1 m</td>
<td>Zone 6</td>
<td>chalcocite-idaite</td>
<td>549.0 ± 11.2</td>
<td>-38.3</td>
</tr>
<tr>
<td>HA-1049-D, 1085.0 m</td>
<td>Zone 5</td>
<td>arsenopyrite</td>
<td>~630 ± 8 to ~670 ± 9**</td>
<td>-18.3</td>
</tr>
<tr>
<td>GDRD1127, 279.9 m†</td>
<td>Zeta</td>
<td>bornite</td>
<td>1012 ± 17</td>
<td>-29.9</td>
</tr>
<tr>
<td>GDRD1127, 262.0 m†</td>
<td>Zeta</td>
<td>chalcopyrite</td>
<td>914 ± 4</td>
<td>-9.6</td>
</tr>
<tr>
<td>PSRD1188, 460.0 m†</td>
<td>Plutus</td>
<td>chalcopyrite</td>
<td>469 ± 27</td>
<td>-19.3</td>
</tr>
</tbody>
</table>

*Sulfur isotopic analysis carried out by the A.I.R.I.E. Program, Colorado State University
**See Table 4.4 for actual results
†Reported in Hall (2013), sulfur isotopic analyses carried out at Colorado School of Mines Stable Isotope laboratory

4.6.3 Xenotime U-Th-Pb geochronology

The EPMA results (Table 4.6) indicated that the xenotime grain in sample MGDD2408_190 is enriched in HREE and MREE relative to LREE with the exceptions of Eu and Lu (Table 4.6). Uranium and lead were present in trace amounts (0.12 to 0.26 wt.%; .01 to 0.5 wt.%, respectively), making the sample a good candidate for U-Th-Pb geochronology. Four successful spot analyses resulted in data (Table 4.7) that form a linear array between an upper, common Pb intercept of \(^{207}\text{Pb}/^{206}\text{Pb}\) of 4833 ± 82 Ma, and a lower intercept of 537 ± 11 Ma (MSWD = 1.5; Figure 4.9a, dashed ellipses). To test the reliability of the data, the integrated data for each spot analysis was split into four individual analyses to obtain more spread along a discordia chord from the xenotime age (lower intercept) to common Pb (upper intercept; Table 4.7; Figure 4.9a, colored ellipses). After discarding three of the data points (XNT_1-2, XNT_1-3, and XNT_3-2) with ~600 to ~700 Ma \(^{207}\text{Pb}\)-corrected \(^{206}\text{Pb}/^{238}\text{U}\) ages that plotted outside of the linear array, the data points yielded an upper, common Pb intercept of 4992 ± 52 Ma, and a lower intercept of 538.4 ± 8.3 Ma (MSWD = 0.74), which is indistinguishable from the age obtained by utilizing the bulk integrated data and is taken as the age of xenotime crystallization.

LA-ICPMS spot analyses from fifteen grains yielded mostly discordant data indicating both high common Pb content and possible radiogenic Pb loss or minor (re)crystallization (Table 4.7). Three concordant analyses have corresponding \(^{207}\text{Pb}\)-corrected \(^{206}\text{Pb}/^{238}\text{U}\) ages of 956 ± 21, 953 ± 24, and 924 ± 21 Ma. The discordant data yielded \(^{207}\text{Pb}\)-corrected \(^{206}\text{Pb}/^{238}\text{U}\) dates between 952 ±26 Ma and 706 ± 27 Ma. A Discordia chord could not be fitted to the data due to the high common Pb contents of the grains. At best, the data gives a strong indication that the xenotime initially formed at ~950 Ma and was affected by subsequent events that resulted in disturbance of the isotopic system.
Table 4.6 Electron microprobe analyses of xenotime, sample MGDD2408_190, Northeast Mango 2 Prospect, Kalahari Copperbelt, Botswana.

<table>
<thead>
<tr>
<th>Element</th>
<th>Elemental Weight Percent (wt. %)</th>
<th>Oxide</th>
<th>Oxide Weight Percent (wt. %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample MGDD2408_190b</td>
<td>Spot 1</td>
<td>Spot 2</td>
<td>Spot 3</td>
</tr>
<tr>
<td>P</td>
<td>15.83</td>
<td>15.9146</td>
<td>15.7865</td>
</tr>
<tr>
<td>O</td>
<td>32.4641</td>
<td>32.6215</td>
<td>32.405</td>
</tr>
<tr>
<td>Sc</td>
<td>0.016801</td>
<td>0.020493</td>
<td>0.017746</td>
</tr>
<tr>
<td>Y</td>
<td>36.4095</td>
<td>36.9953</td>
<td>36.9249</td>
</tr>
<tr>
<td>Ce</td>
<td>0.003843</td>
<td>0.000458</td>
<td>0.001051</td>
</tr>
<tr>
<td>Pr</td>
<td>0.028826</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Nd</td>
<td>0.094898</td>
<td>0.040188</td>
<td>0.038449</td>
</tr>
<tr>
<td>Sm</td>
<td>0.37473</td>
<td>0.267626</td>
<td>0.260809</td>
</tr>
<tr>
<td>Eu</td>
<td>0.06853</td>
<td>0.051981</td>
<td>0.029623</td>
</tr>
<tr>
<td>Gd</td>
<td>2.95632</td>
<td>2.45645</td>
<td>2.23924</td>
</tr>
<tr>
<td>Tb</td>
<td>0.717916</td>
<td>0.667334</td>
<td>0.687794</td>
</tr>
<tr>
<td>Dy</td>
<td>5.89837</td>
<td>5.7905</td>
<td>5.64969</td>
</tr>
<tr>
<td>Ho</td>
<td>1.19298</td>
<td>1.13347</td>
<td>1.16321</td>
</tr>
<tr>
<td>Er</td>
<td>2.07872</td>
<td>2.24313</td>
<td>2.32249</td>
</tr>
<tr>
<td>Tm</td>
<td>0.220852</td>
<td>0.229931</td>
<td>0.234439</td>
</tr>
<tr>
<td>Yb</td>
<td>0.813986</td>
<td>0.760435</td>
<td>0.842373</td>
</tr>
<tr>
<td>U</td>
<td>0.122867</td>
<td>0.260718</td>
<td>0.133103</td>
</tr>
<tr>
<td>Pb</td>
<td>0.04073</td>
<td>0.010435</td>
<td>0.051955</td>
</tr>
<tr>
<td>Ca</td>
<td>0.003578</td>
<td>0.010834</td>
<td>0.01091</td>
</tr>
<tr>
<td>Si</td>
<td>0.020509</td>
<td>0.01549</td>
<td>0.026174</td>
</tr>
<tr>
<td>Total</td>
<td>99.3581</td>
<td>99.4909</td>
<td>98.8264</td>
</tr>
</tbody>
</table>

Figure 4.9 Terra-Wasserburg concordia diagrams for xenotime LASS ICP-MS analyses. A) Terra-Wasserburg concordia diagrams for hydrothermal xenotime, Northeast Mango 2 prospect (MGDD2408_190). The diagram displays split time integrations of the four spot analyses (n=13 of 16 accepted analyses). Inset shows cluster of analyses and the lower intercept of the discordia chord. B) Analyses of xenotime grains, Zone 6 prospect (HA-614-D_505.1). A discordia chord could not be fit to the data due to high concentrations of common Pb and possibly radiogenic Pb loss.
Table 4.7 Xenotime LA-ICPMS U-Th-Pb results for the Kalahari Copperbelt, Botswana

<table>
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<th>Spot #</th>
<th>U ppm</th>
<th>Th ppm</th>
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<th>238U/204Pb 2σ</th>
<th>207Pb/204Pb 2σ</th>
<th>232Th/238U 2σ</th>
<th>208Pb/204Pb</th>
<th>205Pb-cor. 238U/232Th age 2σ</th>
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</tbody>
</table>

Sample MGDD2408_190, xenotime inclusion in vein-hosted chalcopyrite, four spots, each as an entire integration

Sample MGDD2408_190, xenotime inclusion in vein-hosted chalcopyrite, four spots, with integrations split to maximize spread

Sample HA-614-D_505.1, xenotime grains adjacent to bedding-parallel Qtz-Cal-Ce-IId-Bn-Dg-native Ag vein
4.7 Discussion

The new Re-Os sulfide ages of 549.0 ± 11.2 and 515.9 ± 2.4 Ma and U-Th-Pb xenotime age of 538.4 ± 8.3 Ma reported here from sulfide-bearing veins of the Kalahari Copperbelt indicate that hydrothermal copper mineralization was broadly synchronous with a prolonged period of basin inversion during the ~600 to ~480 Ma Damara orogeny. The morphology and isotopically light sulfur stable isotopic signature of the finely disseminated molybdenite suggests that diagenetic framboidal pyrite was replaced by molybdenite during hydrothermal copper mineralization. This suggests that in-situ diagenetic sulfides acted as a chemical reductant and facilitated hydrothermal copper sulfide precipitation. Organic material is also a potential means of sulfide precipitation in the belt. Copper sulfides with Pd-Hg tellurides, molybdenite, and emplectite (CuBiS$_2$) were observed within an organic matter-bearing vein from the Quirinus/Boseto West prospect (see Figure 4.1; Piestrzynski et al., 2016).

4.7.1 Mineralizing events related to the Damara orogeny

The U-Th-Pb date of 538.4 ± 8.3 Ma from a xenotime inclusion in vein-hosted chalcopyrite predates and/or overlaps by five to ten million years with the ~530 to ~535 Ma peak metamorphism in the southern foreland zone of the Damara Belt (Ahrendt et al., 1978; Gray et al., 2008; Miller, 2008). The xenotime age indicates vein emplacement during the main phase of regional northwest to southeast-shortening (Figure 4.10). The Re-Os ages of 549.0 ± 11.2 Ma (vein-hosted chalcocite-idaite, Zone 6), 515.9 ± 2.4 Ma (molybdenite, Northeast Fold), and 496 to 422 Ma (vein-hosted chalcopyrite, Plutus deposit; Hall, 2013a) suggests that punctuated or pulsed mineralizing events occurred during progressive deformation in the Ghanzi-Chobe zone (Figure 4.10).

At the low greenschist metamorphic grade conditions operative during deformation, strain partitioning occurred in the Ghanzi Group. Shearing as a result of flexural slip was restricted largely to boundaries between weak and strong lithologies, such as the Ngwako Pan – D’Kar contact (Davies, 2013). The laminated to thinly bedded, commonly carbonate-rich rocks of the D’Kar Formation facilitated brittle-ductile shearing and allowed focusing of hydrothermal fluid flow within the finer-grained stratigraphic intervals. Repeated veining events are indicated by textures resulting from reopening/cracking, brecciation, and shearing of previously formed veins. The new age data indicates that mineralization also occurred repeatedly during basin inversion.

The Re-Os ages for the Kalahari Copperbelt are broadly contemporaneous with ages for base and/or precious metals mineralization in the Damara Belt of central and northern Namibia and the Central African Copperbelt of the Democratic Republic of the Congo and Zambia. In the Damara Belt, molybdenite associated with quartz-bismuth-pyrite-pyrrhotite veins from the Navachab orogenic gold deposit yielded ages of 525-520 Ma (Steven et al., 2014) while molybdenite from the Onganja copper deposit yielded ages of ~517-518 Ma (Moore, 2011). In the Central African Copperbelt, Re-Os data from...
stratiform ores of the Nkana-Mindola, Chibuluma West, and Nchanga deposits indicate sulfide precipitation at ~576 Ma, corresponding to the earliest stages of basin inversion (Selley et al., 2005). Re-Os (molybdenite) and U-Pb (monazite) geochronology from the vein-hosted Kansanshi Cu-Mo-U deposit yielded two distinct ages of mineralization at ~512 and ~502 Ma, indicating a post peak-metamorphic age of sulfide deposition (Torrealday et al., 2000). Re-Os ages of ~450 Ma indicate post-orogenic mineralization at the Kipushi Cu-Zn-Ge ore deposit (Schneider et al., 2007). Thus, the available Re-Os age data suggest that widespread migration of metalliferous hydrothermal fluids occurred in distinct pulses throughout the Kalahari Copperbelt, Damara Belt, and Central African Copperbelt during Pan-African deformation.

### 4.7.2 Pre-Damara mineralizing events in the Kalahari Copperbelt

The pre-deformation ages of molybdenite (981 ± 3 Ma) and marginal LLHR chalcopyrite (~995 to ~1017 Ma; Hall, 2013) in the Ghanzi Ridge area places a minimum age of deposition for the lower D’Kar Formation at approximately 980 to 1000 Ma (Figure 4.11). Recent U-Pb ages for detrital zircons
in the D’Kar Formation indicate a maximum depositional age of approximately 1050 Ma (Figure 4.11; Talavera, 2012; Gill, 2013; Chapter 3). These new geochronology data indicate that the Ghanzi Group was deposited in a late Meso- to earliest Neoproterozoic basin.

The 981 ± 3 Ma molybdenite Re-Os age overlaps with a 981 ± 43 Ma Rb-Sr whole-rock errorchron age obtained from the Goha Hills (Kgwebe Formation equivalent in northeastern Botswana, Figure 11; Key and Rundle, 1981). Plagioclase 40Ar/39Ar plateau ages from Proterozoic dykes within the Karoo (Jurassic) giant Okavango Dyke Swarm yielded a possibly geologically significant weighted-mean
age of 982.7 ± 4 Ma (Figure 4.11; LeGall et al., 2002; Jourdan et al., 2004). These three ages suggest that a regional thermal event that affected the northern margin Kalahari Craton at approximately 980 Ma, a time that coincides with a marine incursion into the Ghanzi basin that may represent a period of extension and subsidence. Regionally, this age is broadly similar to the age of cooling of the high-grade rocks in the Namaqua Belt on the western margin of the Kalahari Craton, marking the final amalgamation of Rodinia (Miller, 2012).

The two ~950 Ma and one ~925 Ma concordant U-Th-Pb xenotime ages suggests that the xenotime grains initially formed at ~950 Ma and represent an authigenic mineral phase that grew within the host rocks during diagenesis (Figure 4.11). Discordance in many of the analyzed grains suggests that they were affected by later events that resulted in isotopic disturbance of the U-Pb system. Although a lower intercept age for these grains could not be calculated due to the presence of common Pb, isotopic disturbance is likely to have occurred during the hydrothermal event that emplaced the adjacent 549.0 ± 11.2 Ma quartz-calcite-chalcocite-idaite-bornite-digenite-native silver vein.

The Re-Os age range of 670 to 630 Ma obtained for the Zone 5 arsenopyrite predates Damara orogenesis by ca. 50 to 100 million years. The size and euhedral character of the arsenopyrite is similar to late diagenetic cubic pyrite that is common in many silty to sandy mudstones from around the world (Elmore et al., 2016). Similar cubic pyrite is present throughout the stratigraphic sequence in the Ghanzi Ridge area and in Namibia. The Re-Os age for arsenopyrite is broadly similar to the whole-rock Pb age of ~600 Ma obtained from the Klein Aub mine in central Namibia (Walraven and Borg, 1992). The Zone 5 arsenopyrite is overprinted by a later phase of Cu-Pb-Zn-(Ni-Co-LREE) mineralization associated with carbonate-quartz veining and cleavage development. Mineralogical associations (Cu-Pb-Zn-Ni-Co-LREE), Re-Os ages, and U-Pb ages observed in other (similar) syn-kinematic hydrothermal veins throughout the Kalahari Copperbelt suggest that the base metal phase that overprints the arsenopyrite was related to the 560 to 490 Ma Damara orogenic event. This follows observations of coarse cubic pyrite that are replaced by copper sulfide minerals at Klein Aub (Borg and Maiden, 1989) and Boseto (Schwartz et al., 1995; Hall, 2013a).

4.7.3 Isotopic inheritance in the Re-Os system: the role of diagenetic sulfides and organic matter

The 981 ± 3 Ma molybdenite from Northeast Mango 2 as well as the 1012 ± 17 Ma bornite and 914 ± 4 Ma chalcopyrite ages obtained from Zeta deposit (Hall, 2013a) pre-date basin inversion by 450 to 500-million-years. This suggests that while the majority of sulfides in the Kalahari Copperbelt are directly related to veins associated with the Damara orogeny, earlier episodes of mineralization did occur.

The textures associated with the 981 ± 3 Ma molybdenite sampled from Northeast Mango Two deposit are strongly suggestive of replacement of pre-existing bacteriogenic sulfide (framboidal pyrite) and or microscopic organic-rich (bacterial?) allochems within the mudstone bed. The -8.2‰ sulfur
isotopic composition of this molybdenite is also compatible with inheritance of biogenic reduced sulfur. Trace amounts of molybdenite replacing disseminated, possibly diagenetic pyrite was observed in the 1012 ± 17 Ma vein-hosted bornite sample from the Zeta deposit (Hall, 2013a). Isotopic studies of sulfide and sulfate minerals from throughout the Kalahari Copperbelt indicate that the contained sulfur is generally light with $\delta^{34}S$ values ranging from +5 to -45‰ (Ruxton, 1981; Hall, 2013a, Morgan, 2013, Shepard, 2014; Walsh, 2014; Piestrzynski et al., 2016) suggesting that much of the reduced sulfur in the Kalahari Copperbelt sulfides may have been derived through bacteriogenic processes. The hydrothermal mineralogical associations, sulfur isotopic composition, and the geochronology data suggest that hydrothermal processes that affected the mudstone clast resulted in recrystallization and replacement of in-situ framboidal pyrite by molybdenite and other sulfide minerals, with the Re-Os systematics of the original diagenetic pyrite being preserved during the process.

Hall (2013a) utilized automated mineralogy analysis from the Plutus deposit indicated at least two generations of mineralogically distinct bedding-parallel veinlets/fractures: early quartz-albite-chlorite-sulfide that are cut by calcite-dolomite-quartz-sulfide with potassic alteration selvages that been shown to be associated with brittle-ductile deformation. The latter overprints the former in some cases, resulting in brecciation and replacement of the earlier quartz-albite-chlorite-sulfide assemblage (Hall, 2013a). The combined age data and textural relationships suggests that earlier structures or mineral phases affected the mechanical behavior of the rocks during deformation and hydrothermal fluid flow. The concentration of ~950-925 Ma xenotime grains along the margin of the Zone 6 ~549 Ma bedding-parallel quartz-calcite-chalcocite-idaite-bornite-digenite-native silver vein also suggests that bedded authigenic mineral phases may have provided rheological contrasts that aided fracturing and/or slip during subsequent deformation.

Recent investigations into North American shales indicate that diagenetic fractures with brittle fill minerals (quartz, calcite, dolomite) can also influence the mechanical behavior of the rock, acting as planes of weakness that may be reactivated during subsequent pulses of hydraulic fracturing (Elmore et al., 2016). In addition to pyrite, these diagenetic fracture networks may contain kerogens derived from the breakdown of organic material (Elmore et al., 2016). Re-Os geochronology has been utilized to determine the timing of hydrocarbon maturation (oil, bitumen) to help identify organic-rich source rocks and to unravel maturation-migration histories of hydrocarbons within sedimentary basins (Finlay et al., 2011, 2012; Stein and Hannah, 2015). These developments have sparked research on the relationships between hydrocarbons and base metal deposits in sedimentary basins. Recent research suggests that the fluids responsible for many sediment-hosted metallic ore deposits in these basins may have a direct link to multiple phases of production and migration of hydrocarbons (e.g. Holdsworth et al., 2013; Stein and Hannah, 2015). In addition, growth of authigenic monazite nodules has been tied to periods of
hydrocarbon expulsion in Pre-Mesozoic mudrocks (Evans et al., 2002). If the D’Kar and Mamuno formations were deposited prior to ~900 Ma (approximate maximum age of the overlying Nosib Group in Namibia, Figure 4.11), then the rocks at the base of the D’Kar Formation would have been within the ‘oil window’ where hydrocarbon maturation occurs (~2-4 km burial depth). The 914 ± 4 Ma chalcopyrite (Hall, 2013a) may represent epigenetic hydrothermal fluids that exploited diagenetic sulfide- and/or hydrocarbon-bearing fractures.

Piestrzynski et al. (2015) described dark organic matter spatially associated with copper sulfide, molybdenite, Bi-minerals, and Hg-Pd telluride minerals in a late hydrothermal quartz-calcite-sulfide vein for the Quirinus prospect (Figure 4.1). They demonstrated a correlation between increased TOC content and base metal content. This suggests that hydrocarbons may have been important for epigenetic sulfide precipitation (Figure 4.10). The Northeast Fold deposit also contains similar indirect evidence for possible mobile hydrocarbons in the Ghanzi basin. Within the crest of the plunging anticline, normally hematitic (oxidized) Ngwako Pan Formation meta-sandstone is cut by fracture-controlled alteration characterized by iron reduction (hematite to pyrite) and bleaching (Hall, 2013b). These chemically reduced and bleached zones control the distribution of sandstone-hosted (redbed) copper mineralized zones at the Northeast Fold deposit, with ore sulfides occurring only where hydrothermal veins or fractures cross-cut the previously altered rock. The sandstone-hosted copper mineralization at Northeast Fold deposit likely formed through the mixing of residual organic material/hydrocarbons that accumulated in the crest of the anticline and brines that were expelled after peak metamorphism (~515 Ma) in the Ghanzi-Chobe zone.

The relationship between wall rock reduction/bleaching and hydrocarbons is observed in many sedimentary rock-hosted copper deposits, especially in sandstone-hosted/redbed type deposits. In the Chu-Sarysu Basin, Central Kazakhstan, broader zones of iron reduction (bleaching) of sandstones and conglomerates of the red-bed sequence extend over 10 km beyond each of the deposits along E-NE-trending anticlines. The bleached zones and organic residues within them are remnants of former petroleum fluid accumulations trapped by the anticlines (Box et al., 2012). At the Lisbon Valley, USA locality, bleached portions of the Wingate Sandstone also contain increased kaolinite, altered ilmenite and feldspar grains, and traces of oil. These characteristics, combined with spatial relationships indicating bleaching by an immiscible and buoyant reducing fluid, lead to the conclusion that bleaching was caused by hydrocarbon migration along the Cashin fault (MacIntyre et al., 2012).

4.8 Conclusions

Hydrothermal, structurally-controlled quartz-carbonate copper mineralization in the Kalahari Copperbelt was characterized by a metal assemblage consisting of Cu, Fe, Pb, Zn, Ag, Ti, Mo, Bi, Ni, Co, As, (Hg, Pd, Te, Cd, Se, Sb) as well as LREE-bearing phosphates, carbonates, and fluoro-carbonates and
was associated with cryptic potassic alteration. New Re-Os ages indicate that base metal mineralization occurred in at least two main pulses associated with progressive deformation of the Ghanzi Group. U-Th-Pb xenotime (538.4 ± 8.3 Ma) and Re-Os chalcocite-idaite (549.0 ± 11.2 Ma) ages indicate an initial pulse during southeast shortening in the Ghanzi-Chobe zone. The Re-Os molybdenite age of 515 ± 2 Ma post-dates peak metamorphism (~530 Ma) in the Southern Foreland Zone of the Damara belt in Namibia by 15 million years. This suggests a post-peak metamorphic mineralizing event. The timing of multiple pulses of syn-kinematic hydrothermal mineralization in the Kalahari Copperbelt coincides with that of some mineralizing events in the Central African Copperbelt in Zambia and the Democratic Republic of the Congo as well as base- and precious metal mineralization in the Damara Belt in Namibia. The data suggests that hydrothermal mineralizing events on both the Kalahari and Congo cratons may be temporally linked through regional-scale tectonic events.

Based on a 981 ± 3 Ma Re-Os age for molybdenite that replaces diagenetic framboidal pyrite, it is concluded that some of the Re-Os in epigenetic hydrothermal sulfides may have been inherited from both diagenetic iron sulfides and possibly organic matter. These data indicate that the Re-Os systematics of sulfide and/or organic matter may not be reset during in-situ replacement by other sulfide phases, particularly molybdenite. The 981 ± 3 Ma Re-Os molybdenite age provides a new minimum depositional age constraint for the D’Kar Formation host rocks. U-Th-Pb ages of ~950-925 Ma obtained for xenotime grains adjacent to a 549.0 ± 11.2 Ma hydrothermal vein are suggestive of post-depositional diagenetic xenotime growth within the basin. These U-Th-Pb ages post-date sedimentation of the D’Kar Formation and indicate that the xenotime grains are of authigenic origin. The range of Re-Os ages of coarse arsenopyrite indicate a ~630 to 670 Ma mineralizing event in the basin that pre-dates epigenetic copper and base-metal sulfides. Diagenetic mineral phases and/or fractures affected the mechanical behavior of the rocks and enhanced structural permeability during subsequent mineralizing events. The data suggests that the Re-Os systematics of pre-existing, in-situ sulfide minerals and/or organic material was, in-part, retained during replacement by hydrothermal sulfide mineral phases including molybdenite, bornite, and chalcopyrite.

4.9 Acknowledgements

The authors thank the staffs of Khoemacau Copper Mining (Ltd) Pty., Cupric Africa, and Discovery Metals Botswana, not only for their sponsorship of this project, but also for discussions regarding the geology of the Kalahari Copperbelt. We would especially like to thank Mary Toteng, Wallace Mackay, John Deane, and David Catterall. The authors thank Heather Lowers of the USGS and Katharina Pfaff at the Colorado School of Mines for their assistance with microanalytical techniques, Gang Yang and Aaron Zimmerman of the AIRIE Program for the careful analytical work, Craig Johnson and Cayce Gulbransen of the USGS for stable isotope work, and Nigel Kelly for his guidance on
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CHAPTER 5
MAGNETIC LITHOSTRATIGRAPHY FROM HIGH-RESOLUTION AEROMAGNETIC MAPS: A CASE STUDY FROM THE KALAHARI COPPERBELT, BOTSWANA
A modified version of a paper to be submitted to Geophysics

Wesley S. Hall, Yaoguo Li, Murray W. Hitzman, Piret Plink-Björklund, and M. Stephen Enders

5.1 Abstract
Numerical transformations of high-resolution aeromagnetic grids from the extensively concealed Kalahari Copperbelt in Botswana were investigated to determine if the observed intraformational magnetic fabric could be used to map lithostratigraphy and make inferences about stratigraphic and basin architecture for the rift-related and subsequently inverted, tightly folded, and peneplained Mesozoic to Neoproterozoic Ghanzi Group. Comparison of the aeromagnetic maps to well-constrained stratigraphy reveal a strong correlation between sandstone-dominated packages of the copper-bearing D’Kar Formation and (relative) high-intensity anomalies of the second vertical derivate (2nd VD) map fabric. The 2nd VD acts as a high-pass filter, greatly attenuating the regional magnetic field and accentuating the intraformational, strike-parallel, elongate texture, moderate to low amplitude, high-frequency magnetic fabric. The term magnetic lithostratigraphy is defined to describe the technique of identifying stratigraphic boundaries through the changes in the aeromagnetic map textures, which can be used as an indirect mapping tool, similar in sense to seismic stratigraphy.

The magnetic lithostratigraphy interpretation of the Zone 5 –Northeast Mango Two deposit area suggests that the redbeds of the Ngwako Pan Formation filled a sub-basin that is bounded by and pinches out onto a major footwall paleotopographic high to the southwest. Syn-sedimentary faults adjacent to the paleotopographic high are implied by drastic lateral thickness changes in the aeromagnetic fabric of the underlying basal Kuke Formation. Upward warping and truncation of the Ngwako Pan Formation magnetic fabric by that of the overlying D’Kar Formation implies the presence of a previously unrecognized regional unconformity that was associated with paired (relative) uplift and subsidence of the underlying geologic formations prior to basin-wide marine transgression and D’Kar Formation sedimentation. Distinctive along strike, shallow to (relatively) deeper water facies changes at the Zone 5 deposit support the conclusion of a persistent paleotopographic high adjacent to the deposit, which was critical in the development of litho-chemical trap sites for subsequent mineralizing events.

5.2 Introduction
The Cenozoic to recent Kalahari Basin conceals much of the underlying Precambrian basement terranes of southwestern Africa (Figure 5.1). The southern foreland of the Pan-African (~600 to ~480 Ma) Damara Belt in Botswana and Namibia is host to several sedimentary rock-hosted copper-silver
deposits that are collectively termed the Kalahari Copperbelt (Figure 5.1; Borg and Maiden, 1989). The largest collection of known deposits occurs in northwestern Botswana where a basement window in the central Kalahari Basin, the Ghanzi Ridge, exposes a chain of small basement inliers where the unconsolidated sand cover of the Kalahari Group around the inliers is generally less than 30 meters thick (Figure 5.1). Some of the known resources in this area include the Boseto/Ngwako Pan copper deposits (Boseto district; U.S. Steel, 1978; Schwartz et al., 1995; Hall, 2013), the Banana Zone district, the Zone 5, Zone 5 North, and Northeast Mango Two deposits (Northeast district; Van der Heever and Arengi, 2010; Kehmeier, 2014; Catterall et al, 2015), and the Mahumo and T3 deposits (MOD Resources, 2015, 2017; Figure 5.1).

The copper deposits are hosted by deformed Meso- to Neoproterozoic rocks of the Ghanzi Group, a continental to marine rift succession that overlies bimodal volcanic rocks of the 1.1 Ga Kgwebe Formation (Modie, 1996; Schwartz et al., 1996). This volcano-sedimentary basin forms the northern extension of the ~1.1 Ga Koras-Sinclair-Ghanzi rift system, which wraps the western and northern margins of the composite Kalahari Craton (Figure 5.1; Borg, 1988). The Ghanzi and Chobe basins (collectively known as the northwest Botswana rift; Key and Ayres, 2000) were inverted during southeasterly-directed compression related to the Pan-African (~600 to ~480 Ma) Damara Orogen. This resulted in a ~700-km-long and ~150 to ~200-km-long fold belt, the Ghanzi-Chobe zone (Figure 5.1; Hutchins and Reeves, 1980). The fold belt is in part unconformably overlain by syn-orogenic metasedimentary rocks of the Okwa Group as well as sedimentary rocks and mafic lavas of the Karoo Supergroup; the latter are locally preserved in graben structures (Figure 5.1). Deep erosion of the fold belt left a relatively planar surface which exposed the cores of regional-scale, elongate, doubly-plunging folds (20 to 70-km-long, 5 to 10-km-wide folds with amplitudes on the order of 4-6 km; Schwartz et al., 1995; Modie, 2000; Hall, 2013; Lehmann et al., 2015). The peneplained fold belt was later largely covered by Cenozoic to recent Kalahari sedimentary rocks and unconsolidated sands.

Regional aeromagnetic surveys (fixed wing, 250 m line-spacing, 80 m flight height) conducted by the Botswana Geologic Survey have previously been used to delineate the major lithostratigraphic formations within the Ghanzi-Chobe zone (Figure 5.1; Hutchins and Reeves, 1980; Key and Ayres, 2000; Hall, 2013; Lehmann et al., 2015). Irregularly textured strong magnetic intensity fabrics are related to the Kgwebe Formation porphyritic rhyolites. The unconformably overlying Ghanzi Group was characterized by alternating flat textured, low intensity fabrics and elongate textured, strong magnetic intensity fabrics that are related to folded, magnetite-rich sedimentary rocks (Hutchins and Reeves, 1980; Lehmann et al., 2015). These data were used to locate and define the contact between elastic ‘red beds’ of the Ngwako Pan Formation, which displays a flat texture with low intensity, low frequency anomalies, and the
Figure 5.1 The Kalahari Copperbelt in Botswana. A) Map of the Cenozoic to recent Kalahari Basin in southwestern Africa with outline of the predominantly concealed Kalahari Copperbelt. Modified from Haddon (2005). B) Regional Precambrian tectonic framework of southwestern Africa. Abbreviations: CACB – Central African Copperbelt; Ch – Chobe basin; CKB – Choma-Kalomo Block; Gb – Ghanzi basin; GCz/NWBR = Ghanzi-Chobe zone/northwest Botswana rift; KAb – Klein Aub basin Kb – Koras basin; Sb – Sinclair basin; SFz = Southern Foreland Zone of the Damara Orogen; ZB – Zambezi Belt. Modified from Lehmann et al. (2015). C) Inferred geology of the Ghanzi Ridge study area overlain on gray-scale, regional fixed-wing reduced to the pole (RTP) first vertical derivative (1st VD) aeromagnetic grid (250 m line-spacing, 80 m flight height; Geological Survey of Botswana). Inset: stratigraphic column of the Ghanzi-Chobe zone in Botswana, modified after Modie (1996). Abbreviations: KWf = Kgwebe Fm.; NPf = Ngwako Pan Fm.; Df = D’Kar Fm.; Mf = Mamuno Fm.; OKg = Okwa Group.
cupriferous D’Kar Formation, which displays an elongate texture with high intensity, moderate to high frequency anomalies in Total Magnetic Intensity (TMI) datasets (Lehmann et al., 2015). Exploration activity has provided a wealth of data to aid in the geologic interpretation of the fold belt. Extensive exploration drilling and acquisition of high-resolution aeromagnetic (helicopter-borne 75 m line-spacing and 20 m flight height) and ground magnetics datasets within the northeastern sector of the Ghanzi Ridge area provides a unique opportunity to establish relationships between lithostratigraphy and the observed magnetic fabrics. The first and second vertical derivative filters (1st VD and 2nd VD, respectively) are used to enhance low-amplitude, high-frequency anomalies like the Ngwako Pan Formation and suppress high amplitude, long wavelength anomalies emanating from the underlying Kgwebe Formation. The resultant 2nd VD maps are compared to well-constrained deposit-scale lithostratigraphy from the Plutus (part of the Boseto copper deposits) and Zone 5 copper-silver deposits to establish relationships between the 2nd VD magnetic fabric and lithostratigraphy. The resultant magnetic lithostratigraphy maps are used to investigate the stratigraphic architecture of the basin and its implications for basin evolution and models for copper mineralization.

5.3 Data and methods

Aeromagnetic data are relatively cost-effective and quick to acquire compared to other geophysical data sets and allow mapping of geological structure at depth and under cover. The Geological Survey of Botswana has released data from 15 high-resolution aeromagnetic surveys (at 250 m line spacing and 80-m flight height) that cover approximately 90% of Botswana (Appendix B). Between 2010 and 2011, mineral exploration companies conducted three high-resolution, helicopter-borne aeromagnetic surveys (at 75 m line-spacing and 20 m flight height) that covered seven blocks within the Botswana portion of the Kalahari Copperbelt. The geophysical data covering the Ghanzi Ridge, an area with abundant exploration boreholes that allow for lithostratigraphic correlations and petrophysical sampling, were made available for this study.

Total Magnetic Intensity (TMI) maps were interpolated to a spatial grid of 15 meters. Filtered products calculated from the high-resolution magnetic TMI grids were numerically transformed to enhance different characteristics of the magnetic field. Prior to these transforms, a micro-levelling technique combining a Butterworth filter (8th order with a cutoff wavelength of 4 times the line spacing) and a directional cosine filter (with a cut-off azimuth of 0 for the N-S flight lines) was applied to minimize flight line noise. To preserve the spatial resolution of the resulting transformations, the cell size of the TMI grid was first reduced from 15 m to 4 m prior to reduction-to-pole (RTP) operation. The RTP transformation reshapes dipole anomalies by re-calculating the TMI as if the magnetic anomaly is the vertical field produced by vertical magnetization and moves anomalies to better correlate laterally with the causative body (Baranov and Naudy, 1964). The grid was down-sampled back to 15 m prior to
applying additional transformations to the RTP grid. The down-sampling is for numerical efficiency in subsequent processing. Transformed products included analytical signal (AS), tilt derivative (TDR), first and second vertical derivatives (1st VD, 2nd VD, respectively), and total horizontal derivative (THDR) (Miller and Singh, 1994; Cooper and Cowan, 2004; Verduzco et al., 2004).

Because the eroded fold belt exposes several kilometers of stratigraphic section on the limbs of folds, variations in the intraformational magnetic response should be able to be determined by vertical derivatives provided some geologic factors are known. The first vertical derivative of magnetic data measures the change in the vertical component of a laterally variable magnetic field. The resulting anomalies mark the location of contrasts in magnetic susceptibility between two source bodies and hence is often referred to as an edge detection filter. The Fourier transformation of the data also helps to suppress long wavelength anomalies related to the regional magnetic gradient while accentuating moderate to short wavelength, shallow-seated sources. The second vertical derivative measures the rate of the change of the magnetic field, which also varies laterally across non-uniform geologic features. This quantity is very sensitive to the effects of shallow features as well as the effects of noise. It is not possible to undertake any quantitative analyses of 2nd VD maps except to produce characteristic profiles over known geometric forms. The main advantage of 2nd VD maps is to highlight and clarify features spatially (Cooper and Cowan, 2004).

Petrophysical samples were collected for quantitative characterization of the magnetic properties of the rocks at the Zone 5 copper-silver deposit. Anisotropy of magnetic susceptibility (AMS) and the natural remanent magnetization (NRM) of various lithologies were measured in order to calculate the induced intensity of magnetization (J), which can be modeled and compared to the aeromagnetic response of the rocks. The preliminary results of the petrophysical characterization are presented in Appendix C.

5.4 Qualitative analysis

Both the Plutus (Boseto) and Zone 5 Cu-Ag deposits (Figure 5.1) have been drilled to depths of greater than 500 m over strike distances of 4 to 10 km. Both deposits have well constrained lithostratigraphic sections and some available magnetic susceptibility data from drill core to aid in the comparison of lithostratigraphy and aeromagnetic data. The thickness of Kalahari Group cover at the Plutus deposit is minimal, averaging ~ 3 to 6 meters of calcrete and unconsolidated sand while the Zone 5 deposit occurs under 30 to 45 m of similar cover.

5.4.1 The Plutus Cu-Ag deposit

The Plutus Cu-Ag deposit is situated on the northwest limb of the Boseto anticline (Figure 5.1). The host rocks dip 50-55° to the northwest. Approximately 500 to 600 meters of D’Kar Formation stratigraphic section (Figure 5.2; modified from Hall, 2013) was logged from six boreholes that were
Figure 5.2 Stratigraphic long-section (ten times vertical exaggeration) of the Plutus copper-silver deposit. Top: Approximately 500 to 600 meters of lithostratigraphic package is divided into first- and second-order coarsening upward depositional units. Successive second-order units have thinner alternating unit components resulting in closely spaced sandstone units near the tops of first-order units. Bottom: representative core samples of the three main lithological units encountered in the lower D’Kar Formation: algal laminated marlstone, normally graded sandstone-siltstone couplets of the Alternating Unit, and massive bedded subarkose sandstone with thin interbeds of sandstone, siltstone-mudstone units; soft sediment deformation features are common in the sandstones.
spaced 0.5 to 3 km apart, spanning a strike length of ~10 km. The lithostratigraphy of the host rocks were evaluated using cross-sections and a ~10-km-long long-section constructed from drill hole data.

### 5.4.1.1 Stratigraphic analysis

The lower member of the D’Kar Formation is composed of at least four ~50- to 150-meter-thick coarsening upward sedimentary cycles, here referred to as first-order depositional units (Figure 5.2). Each first-order depositional unit contains two to four second-order coarsening upward depositional units ranging in thickness from ~5- to ~50-m. The base of each second-order cycle typically consists of marlstone within the lower half of the stratigraphic section; carbonaceous siltstone occupies the base of each second-order unit within the upper half of the stratigraphic section (Figure 5.2). In each second-order depositional unit, the basal marlstone or carbonaceous siltstone beds are overlain by 15- to 50-meter-thick packages of interbedded siltstone and mudstone, here termed the alternating units; the basal alternating unit is referred to as the Ore Zone unit. The alternating units are gradationally to sharply overlain by 10- to 30-meter-thick packages of sub-arkose sandstone within minor interbedded siltstone-mudstone, here termed the sandstone units.

### 5.4.1.2 Aeromagnetics analysis

The Boseto RTP grid (Figure 5.3a) displays an elliptical, high intensity anomaly centered under the Boseto anticline. Sandstones of the Ngwako Pan Formation, which are known to display a smooth textured, moderate amplitude magnetic fabric in aeromagnetic data (i.e. the green color in the northeastern and southwestern extents of the anticline in Figure 5.3a), are known to occur at and near the surface across the width of the erosional surface of the anticline. The high-intensity elliptical anomaly suggests the presence of deeper-seated, strong magnetic sources, presumably the stratigraphically lower volcanic rocks of the Kgwebe Formation (Figure 5.3b).

The contact with the D’Kar Formation (based on drilling) is represented by a sharp boundary between moderate intensity (Ngwako Pan Formation) and low intensity (D’Kar Formation) in the RTP map (Figure 5.3a-b). The magnetic fabric of the basal D’Kar Formation is characterized by two elongate textured, slightly discontinuous, moderate frequency and intensity, strike-parallel anomalies that dissipate towards the northeast (Figure 5.3a). These anomalies can be traced laterally around the southwestern fold closure of the Boseto anticline where the magnetic character changes to higher intensity anomalies and a third anomaly is present within the section (Figure 5.3a). The lower D’Kar magnetic fabric gives way to the elongate, very high frequency, very high intensity anomalies that characterize the magnetic fabric of the D’Kar Formation in Total Magnetic Intensity (TMI) grids (Figure 5.3b). The anticline is cut by two west-northwest-trending linear anomalies of opposite magnetic polarity that correspond to Karoo dolerite dykes (Figure 5.3a). A parallel zone of smooth to mottled textured, low to moderate magnetic intensity occurs adjacent to the southeastern dyke and cross-cuts the Boseto anticline fold closure (Figure 5.3a).
Figure 5.3 Low- and high-pass aeromagnetic filters (RTP = reduced to the pole; 1st VD = first vertical derivative; 2nd VD = second vertical derivative) applied to the Boseto high-resolution aeromagnetic grid. A-B) RTP map and magnetic profile over the Boseto anticline with cartoon cross-section indicating Kgwebe Formation at depth to give rise to the elliptical shaped, low frequency (deep source), high intensity anomaly. C-D) RTP-1st VD map and magnetic profile. The 1st VD high-pass filter removes some of the regional magnetic field caused by the underlying Kgwebe Formation and accentuates some moderate-frequency anomalies within the Ngwako Pan and D’Kar formations; note the strike-parallel trend of some of the elongate (linear) anomalous features in the Ngwako Pan Formation. E-F) RTP-2nd VD map and magnetic profile. The RTP-2nd VD map minimizes the regional magnetic field caused by the Kgwebe Formation at depth and greatly accentuates the high- to very high frequency (shallow) magnetic anomalies of the Ngwako Pan and D’Kar formations that form elongate (linear), noisy textured anomalies that trend parallel to strike. Note that some cross-cutting structures become more apparent. Black dots in B, D, F are drill holes.
Exploration drilling indicates this feature corresponds to a post-fold extensional graben filled with Karoo sedimentary rocks.

The first vertical derivative (1\textsuperscript{st} VD) of the RTP grid (Figure 5.3c) displays many similarities to the RTP grid, but resolves more features within the Ngwako Pan and D’Kar formations. The 1\textsuperscript{st} VD magnetic fabric of the Ngwako Pan Formation is characterized by an elongate texture with moderate frequency, moderate to high intensity, strike-parallel anomalies (Figure 5.3c-d). The lower section of the Ngwako Pan Formation that occurs near the center of the eroded anticline has higher intensity anomalies compared to the upper sections of the formation that are located on the limbs of the anticline (Figure 5.3c). This high intensity is presumably due to the presence of very high magnetic intensity Kgwebe Formation volcanic rocks in near subsurface. The two, elongate texture, strike-parallel anomalies that characterize the D’Kar Formation in the RTP maps have well defined margins in the 1\textsuperscript{st} VD map while several other linear to discontinuous, high-frequency, moderate- to high-intensity strike-parallel anomalies higher up in the stratigraphic section were observed in the 1\textsuperscript{st} VD map (Figure 5.3c-d).

The second vertical derivative (2\textsuperscript{nd} VD) magnetic fabric of the Ngwako Pan Formation is characterized by abundant elongate textured, strike-parallel, high-frequency, moderate to high intensity anomalies or domains that can be traced around the fold closure to the opposite limb of the anticline, suggesting that elongate texture anomalies represent lithology and not structures (Figure 5.3 e-f). The overlying D’Kar Formation displays a similar magnetic fabric to that of the Ngwako Pan Formation except for the two (to three) high frequency, very high intensity, strike-parallel anomalies near the base of the formation. The stratigraphically higher anomalies were also resolved with more precision than the RTP-1\textsuperscript{st} VD map (Figure 5.3e-f).

All three maps (TMI, RTP-1\textsuperscript{st} VD, and RTP-2\textsuperscript{nd} VD) display a prominent, linear, very high intensity anomaly that passes through the hinge of the southwestern Boseto anticline fold closure, cuts across the magnetic fabric of the Ngwako Pan Formation, and abruptly ends along the Ngwako Pan – D’Kar Formation contact on the southeastern limb of the anticline (Figure 5.3, large arrows). A less prominent linear to slightly curvilinear structure was resolved in the RTP-1\textsuperscript{st} VD and RTP-2\textsuperscript{nd} VD maps. This lineament obliquely cross-cuts the magnetic fabric of the Boseto anticline along a west-southwest trend. The lineament is characterized by destruction of the underlying high intensity magnetic fabric of the Ngwako Pan Formation (Figure 5.3).

The magnetic fabric of the RTP-2\textsuperscript{nd} VD map was compared to the D’Kar Formation stratigraphic long-section from the Plutus deposit area (Figure 5.4). The stratigraphic long-section (two times vertical exaggeration) and the 2\textsuperscript{nd} VD aeromagnetic map display many along-strike similarities across the 10-km-long strike-length of the deposit area; the alternating high and moderate intensity, strike-parallel linear anomalies correlate well with the overall along strike stratigraphic architecture of the D’Kar Formation.
Figure 5.4 Qualitative comparison of the second vertical derivative (2\textsuperscript{nd} VD) map and stratigraphy, Plutus Cu-Ag deposit. A) 2\textsuperscript{nd} VD map (plan view). The 2\textsuperscript{nd} VD aeromagnetic map has been re-oriented (from normal map coordinates) so that the map displays the correct horizontal stratigraphic orientation. The white dashed line delineates the Ngwako Pan-D’Kar Formation contact. Black dots represent locations of inclined drill holes utilized to construct the stratigraphic long-section in (B). Black lines mark the boundaries between low intensity linear anomalies (yellow arrows) and high intensity linear anomalies (white arrows). B) Modified stratigraphic long-section after Hall (2013) depicting the Ngwako Pan-D’Kar contact as the datum, with alternating units (green, yellow arrows, large arrow is the Ore Zone unit) and coarse-grained sandstone units (shaded red, white arrows) components of the first-order coarsening upward units (numbered yellow triangle).
The basal low to moderate magnetic intensity domain of the D’Kar Formation corresponds to the Ore Zone Unit; the RTP-2nd VD map and the stratigraphic long-section both display a decrease in thickness for this unit from southwest to northeast (Figure 5.4). The Ore Zone Unit is overlain by the first prominent, high to very high intensity linear anomaly. This anomaly corresponds to the sandstone-dominated upper half of the basal first-order coarsening upward sedimentary cycle. A faint linear fabric was observed on the map within the prominent linear anomaly. This faint fabric probably resulted from the presence of interbedded fine-grained intervals with the upper portion of the cycle.

The overlying Alternating Unit 1 and Sandstone Unit 2 correspond to well-resolved, low and high magnetic intensity strike-parallel anomalies, respectively. The upper portions of the Sandstone Unit 2 contain thin, discontinuous, low magnetic intensity zones that probably reflect interlayered finer-grained rocks within the upper half of the coarsening upward cycle. The overlying alternating units (2 and 3) and sandstone units (3 and 4) form laterally continuous, elongate anomalies across the deposit area, although they are less well resolved that the underlying units (Figure 5.4). Overall, the qualitative analysis indicates that the laterally continuous high amplitude anomalies correspond to coarser-grained and thicker bedded sandstone-dominated rock packages while the intervening moderate intensity anomalies correspond to the finer-grained rock packages.

5.4.2 The Zone 5 Cu-Ag deposit

The Zone 5 Cu-Ag deposit is located on the southeastern limb of a regional doubly-plunging anticline to dome-like structure located ~10-km to the east-southeast of the Kgwebe Hills (Figure 5.1). The D’Kar Formation host rocks dip approximately 65° to the southeast. The Zone 5 North deposit area occurs 4.5 km to the north on the opposite side of the dome-like anticlinal structure. The host rocks are overlain by approximately 35 to 45 meters of Kalahari Group cover.

5.4.2.1 Stratigraphic analysis

The lower D’Kar Formation at Zone 5 is characterized by alternating packages of relatively fine- and coarse-grained lithologies that overlie medium- to coarse-grained sandstone of the Ngwako Pan Formation (Figure 5.5). The lowermost fine-grained package, termed the Ore Zone Unit, begins with massive to bedded limestone and/or thinly bedded marlstone that passes up stratigraphic section into calcareous siltstone, rhythmically bedded siltstone-mudstone, carbonaceous siltstone, and normally graded sandstone-siltstone-mudstone beds (Figure 5.5).

The Ore Zone Unit is consistently overlain by a ~60- to ~70-meter-thick sandstone-dominated package termed the Marker Sandstone Unit (Figure 5.5). It consists of thick to massive bedded sandstone with intervals of thin to thick interbedded siltstone and sandstone. The Marker Sandstone Unit is in turn overlain by a ~35 to 40-meter-thick Alternating Unit A, which consists of interbedded, thin to thick bedded siltstone, calcareous siltstone to marlstone, and very fine-grained sandstone to sandy siltstone.
Figure 5.5 Generalized lithological measured section from the Zone 5 Cu-Ag deposit and corresponding magnetic susceptibility profile. Note the contrast in magnetic susceptibility between the lower and upper D’Kar Formation.

(Figure 5.5). This unit shares many of the sedimentological characteristics as the Ore Zone Unit, but contains notably less carbonate material. As with the Ore Zone Unit, the Alternating Unit A is overlain by a 70- to 80-meter-thick package of massive sandstone with interbedded zones of medium- to thin-bedded silty sandstone termed the Sandstone Unit A (Figure 5.5). The coupled Alternating and Sandstone units are repeated up to five times in the overlying stratigraphic package (units B, C, D, and E; Figure
However, the successive alternating units become progressively thinner, as thin as ten meters, and more carbonate-rich up stratigraphic section. The sandstone units (B, C, D, and E) retain thickness of ~50-80 meters but become dominated by medium to thin bedded sandstone with thin interbedded siltstones.

The upper member of the D’Kar Formation consists of repetitive three- to fifteen-meter-thick fining-upward packages of brown-red colored planar to cross-stratified sandstone with erosional bases that are transitionally overlain by thin intervals of green-gray mudstone-siltstone. Thin carbonate beds occur sporadically throughout the upper D’Kar Formation and where present generally occur near the top of the fining-upward packages (Figure 5.5).

Chemically reduced lithologies of the lower D’Kar Formation have magnetic susceptibilities that range from 0.5 to 2.0 x 10^{-4} SI units (Figure 5.5); the magnetic susceptibility of sandstones is slightly higher, varying from 3.0 to 5.0 x 10^{-4} SI units. Reduced lithologies in the overlying transition zone have low magnetic susceptibilities, like those of the underlying reduced facies lithologies. However, the interbedded, chemically oxidized sandstones have magnetic susceptibilities that range from 5.0 to 15.0 x 10^{-4} SI units (Figure 5.5). The Upper D’Kar Formation is characterized by thicker packages of oxidized sandstone with magnetic susceptibilities that vary from 5.0 to 40.0 x 10^{-4} SI units while reduced, interlayered fine-grained lithological units are characterized by low susceptibilities (Figure 5.5, see also Appendix C for more information on petrophysical characterization of the Zone 5 lithologies).

### 5.4.2.2 Aeromagnetics analysis

Given the observed correlation between aeromagnetic fabrics and the stratigraphic units (Figure 5.4), analysis of the high-resolution 2nd VD aeromagnetic map over the Zone 5 deposit was undertaken to identify and map the major stratigraphic units. The analysis was also used to map structural elements in the deposit area based on off-sets of magnetic fabrics and/or cross-cutting lineaments that destroy the underlying magnetic fabric.

In regional RTP and 1st VD aeromagnetic maps, the Zone 5 deposit occurs on the southeastern margin of a poorly defined, elliptical- to boudin-shaped (elongate in the regional northeast-southwest trend), moderate intensity TMI anomaly (Figure 5.6a-b). It occurs along strike from a similarly-shaped aeromagnetic anomaly that corresponds to sub-cropping Kgwebe Formation rhyolitic volcanoclastic rocks of the Makgabana Hills (Figure 5.6). The two anomalies are discontinuously linked by a trend-parallel, moderate to high intensity lineament that can be traced along strike in regional aeromagnetic maps for ~85 kilometers along the southwestern flank of the Ghanzi Ridge (Figures 5.1 and 5.6). Other than the underlying long wavelength anomalies, the Ngwako Pan Formation has a smooth textured, low frequency, moderate to low amplitude magnetic fabric in both TMI and RTP datasets. The prominent high frequency, very high amplitude anomalies that characterize the upper D’Kar Formation in both TMI...
Figure 5.6 Zone 5 Cu-Ag deposit aeromagnetics. A) Regional RTP map covering the Zone 5 Cu-Ag deposit and surrounding area, including the Kgwebe and Makgabana Hills (Kgwebe Formation bimodal volcanic rocks) exposed in the cores of anticlines. B) Regional RTP-1st VD map, same area as in (A). Anomalies are better defined due to the suppressed regional magnetic gradient. C) High-resolution aeromagnetic RTP-2nd VD map overlain on the regional RTP-1st VD map covering the Zone 5 orebody (pink polygon). Dashed white line outlines area of well-defined geology by drilling. Solid white box indicates location of the cross-section used to compare lithostratigraphy and aeromagnetic signal. Note the elongate (linear to curvilinear) texture of the high-frequency, alternating high and low intensity 2nd VD anomalies (large black arrows), which trend parallel to the overall strike of the fold limb, suggesting a possible stratigraphic control on the 2nd VD magnetic fabric of the Ghanzi Group.
and RTP maps have a slightly curvilinear character. They do not conform to the elliptical-shape anomalies within the underlying Ngwako Pan and Kgwebe Formation. The Ngwako Pan – D’Kar Formation contact cannot be resolved in the TMI, RTP, and RTP-1st VD maps due to the long wavelength TMI anomaly associated with the probable underlying volcanic rocks (Figure 5.6).

Several drill holes were utilized to determine the subsurface expression of the contact between the Ngwako Pan and D’Kar formations (Figure 5.6), which was then mapped and compared to the filtered aeromagnetic maps covering the deposit area. A representative cross-section of the Zone 5 geology was constructed along a fence of drill holes located in the southeastern lobe of the deposit area that cover up to 1000 meters (true thickness) of D’Kar Formation stratigraphy (Figure 5.6).

Both the total magnetic intensity map and the reduced-to-pole map have smooth textured, high and moderate amplitude anomalies, respectively, across much of the deposit area; the amplitude increases to the southeast approaching the more characteristic elongate texture of the upper D’Kar Formation. Both the TMI and RTP maps failed to resolve the Ngwako Pan-D’Kar formation contact that was determined from drill holes (Figure 5.6). The Ngwako Pan - D’Kar Formation contact was resolved after applying the 1st VD filter, with the combined Ore Zone Unit and overlying Marker Sandstone Unit forming a single linear, positive anomaly within the smooth textured, low frequency, moderate amplitude background of the Ngwako Pan and lower D’Kar formations. A faint linear anomaly above this may correspond to Sandstone Unit A. The elongate, laterally continuous aeromagnetic boundaries of the transition zone (moderate frequency, low intensity anomaly) and the upper D’Kar Formation (moderate-high frequency, high to very high amplitude anomalies) are well-resolved by the 1st VD map (Figure 5.6).

In the second vertical derivative map, the Ngwako Pan Formation has a mottled texture with a noisy curvilinear aeromagnetic fabric defined by moderate frequency, high or moderate amplitude anomalies with at least one reasonably well-defined edge (Figure 5.6). The lower D’Kar Formation has an elongate, high frequency, moderate to high intensity magnetic fabric with strike-parallel linear anomalies; the lower two high amplitude anomalies have well defined edges while the stratigraphically higher, high amplitude anomalies are less well defined and have a slight discontinuous linear fabric (Figure 5.6). The transition zone is marked by a transition from the moderate intensity background of the lower D’Kar Formation to a low intensity background with moderate to low frequency, moderate intensity anomalies (Figure 5.6). The boundary to the upper D’Kar Formation is marked by a shift to a high frequency, very high intensity, linear aeromagnetic fabric (Figure 5.6).

Comparison of the 2nd VD map to a shallow geologic cross-section (Figure 5.7) indicates that the Ore Zone Unit correlates with a moderate intensity linear anomaly with a poorly defined edge (contact) with the Ngwako Pan Formation and a well-defined upper edge that marks the boundary with the overlying Marker Sandstone Unit. The upper and lower margins of the Marker Sandstone Unit correlate
Figure 5.7 Magnetic profiles and aeromagnetic map sections with drill hole collars for TMI, RTP, RTP-1\textsuperscript{st} VD, and RTP-2\textsuperscript{nd} VD and a shallow-level stratigraphic cross-section of the Zone 5 Cu-Ag deposit. The moderate to high intensity linear anomaly in the 1\textsuperscript{st} VD map and profile (yellow-orange, left of image) approximates the contact the between the Ngwako Pan and D’Kar formations. The high amplitude 2\textsuperscript{nd} VD anomalies mark the well- to moderately well-defined lower and upper edges of the Marker Sandstone Unit and Sandstone Units A and B. Only the lower edges of Sandstone Units C-E are marked by moderate to high intensity anomalies with poorly-defined edges. A similar pattern emerges in the Transition Zone, but anomalies are shifted to lower values (low to moderate intensities).

with the peaks of the high intensity linear anomaly and a separated by an intervening discontinuous linear domain of moderate to high intensity; the peaks of the anomalies partially coalesce above the Marker Sandstone Unit in places (Figure 5.7). Sandstone Units A and B have similar aeromagnetic responses to the Marker Sandstone Unit, although the edges of the anomalies are less well defined for each successive sandstone unit; the intervening moderate intensity linear anomalies mark the positions of the Alternating Units A and B (Figure 5.7). Sandstone Units C-E have magnetic anomalies that only occupy the lower
margin of the sandstone units and delineate the contact between underlying carbonate-rich fine-grained rocks and overlying thickly bedded to massive sandstone (Figure 5.7). A similar pattern occurs within the Transition Zone, although the background is shifted to a low intensity anomaly with moderate frequency, moderate intensity linear anomalies that mark the apparent base of sandstone units that overlie alternating units; this is due to the nature of the high-amplitude (and high magnetic susceptibility) of the adjacent (overlying) rocks of the Upper D’Kar Formation (Figure 5.7). The aeromagnetic analysis suggests that the higher intensity linear anomalies mark the lower and sometimes the upper contacts of sandstone units.

5.5 Magnetic lithostratigraphy

In the Ghanzi Ridge area, the regional RTP magnetic field and the 1st VD residual magnetic field are both dominated by the high intensity (~330 to 600 nT), high amplitude magnetic fabrics of the Kgwebe and upper D’Kar formations (Lehmann et al., 2015). However, the 2nd VD aeromagnetic maps greatly attenuate the deep-seated regional magnetic field and accentuate the shallow-seated magnetic sources within the Ghanzi Group. The high-resolution aeromagnetic surveys helped resolve an intraformational, elongate to noisy textured, high-frequency, moderate to high amplitude aeromagnetic fabric for these relatively weakly magnetic metasedimentary rock formations.

5.5.1 The intraformational aeromagnetic signature of the Ghanzi Group

The qualitative comparison of lithostratigraphy and the 2nd VD aeromagnetic maps indicates that the 2nd VD aeromagnetic fabric is strongly controlled by intraformational lithological variations at or near the erosional surface of the dipping strata. The boundaries between elongate, alternating high and low amplitude intraformational anomalies coincide remarkably well with most of the lithostratigraphic contacts of the D’Kar Formation, with high intensity anomalies that occur over the margins of coarser-grained sandstone units and low intensity anomalies that coincide with fine-grained siliciclastic and carbonate rock units. The boundaries to lithostratigraphic units that are between ~60 to 100-meters-thick were well resolved in the 2nd VD map, often with sharp edges. Thinner lithostratigraphic units down to ~20-30 meters were resolved with less confidence due to noise, but a high intensity anomaly was found to generally correspond to instances where a sandstone unit overlies a thin fine-grained unit.

A one order of magnitude break in magnetic susceptibility was previously reported between the oxidized, hematite-stable rocks of the upper Ngwako Pan Formation and the reduced, magnetite-stable lowermost D’Kar Formation (Lehmann et al., 2015). The increase in magnetic susceptibility was due to the presence of minor disseminated monoclinic pyrrhotite and/or magnetite (chemically reduced) within the magnetic fraction of the mineralized basal D’Kar Formation (compared to the chemically oxidized, hematite-bearing upper Ngwako Pan Formation). Some undeformed sandstone beds that occur within the D’Kar Formation contain minor disseminated magnetite that can be weakly detected with a hand magnet. At some localities, magnetite is concentrated along heavy mineral laminations in swaley cross-beds,
indicating that some of the contained magnetite is of detrital origin. Authigenic magnetite phases have
been described previously both in sandstones and in the fine-grained lithological units (Borg, 1988;
Schwartz et al., 1995; Hall, 2013). Detrital magnetite in sandstone beds is the most likely underlying
cause of the stronger magnetic intensity displayed by the sandstone units, although further petrophysical
work is in progress to confirm this mineralogical relationship.

The well-defined margins of 2nd VD aeromagnetic fabrics was used to extrapolate the D’Kar
Formation intraformational magnetic fabric along strike in both directions to the northeast of the Zone 5
deposit area and extending ~12 km along strike to the southwest to the Northeast Mango Two Cu-Ag
deposit (Figure 5.8). The stratigraphy at Northeast Mango 2 is remarkably similar to that of Zone 5 based
on deep exploration drill holes. When compared to the extrapolated intraformational magnetic fabric, the
aeromagnetic response of the alternating and sandstone lithostratigraphic units at Northeast Mango Two
is nearly identical to that of Zone 5, confirming the correlation between lithostratigraphic units and their
associated aeromagnetic fabric. The term magnetic lithostratigraphy is defined to refer to the technique of
correlating lithostratigraphic units with their associated aeromagnetic fabric for mapping purposes, a
process similar in sense to seismic stratigraphy.

The magnetic lithostratigraphy of the D’Kar Formation can be confidently mapped along fold
limbs across large distances (>10 km) using a combination of high-resolution aeromagnetic 2nd VD maps
and geologic data from deep stratigraphic boreholes. Additionally, the 2nd VD aeromagnetic response of
the under- and overlying Ngwako Pan and Mamuno formations, for which there is very little to no
available geologic data, behave in a similar fashion to that of the D’Kar Formation and can be interpreted
in a similar fashion using magnetic lithostratigraphy (Figure 5.8). This powerful indirect mapping tool
can aid with structural mapping and interpretations of the stratigraphic and basin architecture.

5.5.2 Structural mapping with the 2nd VD map

Magnetic lithostratigraphy also aids in defining the position of major structures as well as allows
for the recognition of previously unrecognized structures. The margins of the cross-cutting Karoo
dolerite dykes were resolved with detail while the extent of the fabric destructive alteration associated
with the dykes was also resolved. The folded magnetic lithostratigraphic units can be used to confidently
mark the axial traces of the major anticlines and synclines in the area. Smaller-scale folds, such as
parasitic folds on fold limbs, can also be resolved with high confidence by the 2nd VD map. Offset of
underlying magnetic lithostratigraphic units by both fabric destructive and non-destructive lineaments
were interpreted as faults and their relative displacement in map view can be inferred.

Two dominant structural trends were recognized in the magnetic lithostratigraphy mapping
exercise. The first is a north-northeast-trending zone of structures that occur on the northeastern side of
the Zone 5 deposit (to the left in Figure 5.8). These structures also occur on the southwestern side of the
Figure 5.8 Magnetic lithostratigraphy of the Zone 5 and Northeast Mango Two Cu-Ag deposit areas. The 2nd VD aeromagnetic map (top) has been re-oriented (from normal map coordinates) so that the map displays the correct horizontal stratigraphic orientation. The contacts between the lowermost units of the D’Kar Formation can be confidently mapped based on the second vertical derivative dataset. Note that the units comprising the lowermost D’Kar Formation, including the Ore Zone Unit, are inferred to pinch towards the northeast and form a thin drape over the underlying Zone 5 dome; the transition zone and upper D’Kar Formation overstep the dome without significant variations in thickness. The uppermost part of the map occurs on the opposite limb of a faulted doubly plunging syncline and is not included in the interpretation. Interpretation of geologic units in the Ngwako Pan Formation suggests that the uppermost Ngwako Pan Formation also pinches towards the northeast, suggesting that the Zone 5 dome formed a paleotopographic high.
Mango NE 2 deposit (to the right in Figure 5.8). Some of these structures display right-lateral offset in map view, on the order of a few hundred meters in places. A set of west-northwest-trending structures parallel, in part, the cross-cutting Karoo dolerite dykes near Northeast Mango Two (Figure 5.8). Structures with this orientation also occurs on the eastern side of the Zone 5 deposit, without associated Karoo dolerite dykes (Figure 5.8). Many of the inferred faults are concentrated near the two deposit areas (Figure 5.8).

The west-northwest-trending structures are truncated at or near the Ngwako Pan-D’Kar Formation contact and do not continue into the overlying D’Kar Formation. This could suggest a that the contact itself represents a major structure that halted the propagation of the faults, or that the faults represent an early phase syn-sedimentary faulting that only affected the pre- to syn-rift rock packages and not the overlying post-rift sag succession. This is hinted at in the 2nd VD map where the abrupt thickness changes in the Kuke Formation (the mini-basin that occurs below the Northeast Mango Two deposit area) are inferred to coincide with a pair of west-northwest-trending faults that bound the mini-basin. Although the north-northeast-trending structures display a clear cross-cutting relationship with the folded stratigraphic package, it is plausible that these faults were localized along reactivated basement structures.

5.6 Stratigraphic architecture and implications for basin evolution

Despite previous attempts to characterize the sedimentological evolution of the Ghanzi Group depositional sequence and the tectonic setting of the basin (e.g. Modie, 1996, 2000; Kampunzu et al., 1998; 2000), the sparse availability of geologic data precluded detailed, continuous along-strike correlations of the stratigraphic package. The magnetic lithostratigraphy interpretation of the 2nd VD map over the Zone 5 and Northeast Mango Two deposit area (Figure 5.8) highlights several previously unrecognized stratigraphic relationships. This includes important architectural details that can give insight into the evolution of the volcano-sedimentary Ghanzi basin.

In regional aeromagnetic TMI and RTP maps, the informal Kuke Formation is difficult to distinguish from the overlying Ngwako Pan Formation. The 2nd VD map reveals a distinctive fabric for this formation with well-defined, moderate frequency and high intensity anomalies that are stratigraphically sub-parallel to one another except for the area adjacent to the Makgabana Hills dome. There, the basal units of the Kuke Formation display significant lateral thickness changes over what is inferred to be a set of small convex-shaped (ridges) and concave-shaped (depressions) features. The coincidence of the ridges and depressions and apparent thickness variations of the Kuke Formation in the 2nd VD map are suggestive of the presence of a syn-sedimentary faults. The cause of the high intensity, high frequency aeromagnetic signal is likely due to sedimentary layers with detritus derived from footwall Kgwebe Formation volcanic rocks that are interlayered with extrabasinal sandstones or fine-grained rocks.
with (relatively) low magnetic fraction content (e.g. Modie, 1996, 2000). The slightly higher magnetic intensity area over the center of the Zone 5 dome suggests that the Kuke Formation occurs in the shallow subsurface while the regional high amplitude anomaly coincident with the Zone 5 dome suggests that Kgwebe Formation volcanic rocks occur at slightly greater depths. The Kuke Formation was likely deposited into a small sub-basin near the fault scarp, probably as an alluvial fan.

The magnetic lithostratigraphy of the overlying Ngwako Pan Formation was mapped using the 2nd VD map (Figure 5.8). The magnetic lithostratigraphic package of the Ngwako Pan Formation forms a curvilinear, alternating concave-up-down-up pattern compared to the relatively flat pattern on the overlying D’Kar Formation. We refer to the concave up areas as the Zone 5 dome (adjacent to the Zone 5 deposit in Figure 5.8) and the Makgabana Hills dome (to the southwest of the Northeast Mango Two deposit, not pictured in Figure 5.8). The magnetic lithostratigraphic units thickening towards the center of the inferred basin between the Makgabana Hills and Zone 5 domes (Figure 5.8). The uppermost magnetic lithostratigraphic units of the Ngwako Pan Formation thin and pinch-out over the domes and the inferred basin margins (Figure 5.8). The patterns revealed by the magnetic lithostratigraphy also suggests that the center of the intervening ~10-km-wide sub-basin was actively subsiding during deposition of these sedimentary rocks. The uppermost magnetic lithostratigraphic units of the Ngwako Pan Formation are warped upward adjacent to the basin margins. This suggests that the margins of the sub-basin are inferred to have undergone (relative) uplift. Truncation of the upward warped magnetic lithostratigraphic units by the overriding magnetic lithostratigraphic units of the D’Kar Formation indicate that these units were eroded above the Makgabana Hills and Zone 5 domes prior to marine incursion. This pattern is inferred as a widespread unconformity occurs at the Ngwako Pan – D’Kar Formation contact. This is also suggested by the overstepping of the basal Ore Zone Unit onto at least two stratigraphically separate units of the Ngwako Pan Formation on the southwest margin of the Zone 5 dome (Figure 5.8).

Along strike to the southwest of the Northeast Mango Two deposit, the D’Kar Formation comes close to being in direct contact with the Kgwebe Formation. The lowermost units of the D’Kar Formation have similar but subdued characteristics to the Ngwako Pan Formation. The magnetic lithostratigraphic units form thicker sequences in the center of the basin and thin onto the domed regions (Figure 5.8). This subtle sub-basin is confirmed by the lateral changes in the depositional environment of the Ore Zone unit. A wedge a limestone that pinches to the northeast and thickens to the southwest is present across the northeast and central lobes of the Zone 5 deposit (Figure 5.9). To the southwest, the lithology changes abruptly into a thick sequence of interbedded siltstone and marlstone. The lateral shift in depositional facies suggests that the limestone wedge was a reef-like structure that built-up around the uplifted Zone 5 dome. The distal marlstone facies were likely formed through erosion of the reef and deposition in the center of the sub-basin by sediment gravity flows coming off the carbonate reef (i.e. carbonate turbidites).
Figure 5.9 Cartoon cross-section of the Zone 5 deposit depicting paired uplift and subsidence during the rift-climax producing a concave down-shaped basin, and lateral facies changes of the basal Ore Zone member carbonates, with a basin-ward thickening wedge of massive to stromatolite-bearing limestone that abruptly transitions to interbedded marlstone and siltstone towards the center of the basin, suggesting a fringing reef structure around the Zone 5 dome.

Similar stratigraphic variations occur at the Northeast Mango Two deposit. These stratigraphic relationships suggest that the lowermost portion of the D’Kar Formation was deposited near the end of the period of maximum subsidence in the basin that resulted in marine inundation (e.g. Prosser, 1993; Gawthorpe et al., 1997; Gawthorpe and Leeder, 2000).

The overlying stratigraphy of the D’Kar (lower, transition, and upper) and Mamuno formations form a continuous, sheet-like to layer-cake magnetic lithostratigraphic succession with only minor lateral variations observed (Figure 5.8). The pattern suggests that these units were deposited in a broad, open basin setting. The laterally continuous, stacked, alternating packages of fine-and coarse-grained units of the lower D’Kar Formation could be explained by several different and/or a combination of processes. The cyclic sedimentation could have resulted from: 1) pulses of basin-wide subsidence that trapped coarse-grained sediments at the basin margins followed by progradation of coarse-grained sediment during periods of quiescence, 2) changes in sediment supply to the basin due to fault movements, or 3) autogenic processes such as deltaic avulsion and related progradational episodes. Overall, the D’Kar Formation lithostratigraphy displays a gradual coarsening and shoaling upward trend reflecting the gradual shift from moderate to shallow marine (lower D’Kar) to shallow to emergent (transition to upper D’Kar) to near-shore to fluvial-deltaic conditions (Mamuno Formation). The characteristic of the upper half of the Ghanzi Group suggests that these sediments were deposited during the post-rift sag basin phase of basin development.
5.7 Implications for sedimentary rock-hosted copper mineralization

The Zone 5 and Northeast Mango Two Cu-Ag deposits occupy positions within the basin that coincide with apparent uplifted, dome-like features. Syn-sedimentary faults were probably responsible for the paired uplift and subsidence observed in the basin architecture. These zones of uplift created a sub-basin architecture within the underlying Kuke and Ngwako Pan formations that was overlain by a seal composed of fine-grained siliciclastics and carbonates. The underlying basin architecture was probably conducive to long-lived circulation of basinal brines capable of scavenging metals from the Kgwebe Formation volcanic rocks and the oxidized red beds of the Kuke and Ngwako Pan formations.

The oblique cross-sectional view of the Zone 5 – Northeast Mango Two sub-basin provided by the 2nd VD map (Figure 5.8) indicates that the sub-basin is ~12 km wide, is filled by pre-rift, footwall-derived volcanic rocks and sandstones (Kuke Formation) that are overlain by thick sequence of monotonous sandstones (Ngwako Pan Formation), and is in turn capped by a basin-wide seal consisting of reduced facies fine-grained siliciclastic rocks, limestones, and marlstones deposited in a shallow, storm-dominated shelf environment (D’Kar Formation). The two deposits occur near the margins of the sub-basin adjacent to zones of significant lateral thickness changes within the underlying Ngwako Pan and Kuke formations. Additionally, the Ngwako Pan Formation forms a concave down, bowl-like shape with the uppermost stratigraphic units thinning and pinching towards the margins of the basin where one might expect to find syn-sedimentary faults that bound uplifted footwall blocks. Figure 5.10 shows that the sub-basin architecture shares many similarities to other classic examples of sedimentary rock-hosted stratiform copper deposits from districts around the world including the Polish Kupferschiefer and some deposits of the Zambian Copperbelt (e.g. Hitzman et al., 2005; Selley et al., 2005; Hitzman et al., 2012).

The underlying basin-architecture illustrates the strong link between the inferred locations of syn-sedimentary faults and the locations of stratiform ore deposits. The magnetic lithostratigraphic mapping method developed here can be applied to other districts in the Kalahari Copperbelt to reveal for similar patterns in the under- and overlying stratigraphic architecture, which can aid in exploration targeting throughout the several thousands of line-kilometers of tectonically repeated Ngwako Pan – D’Kar Formation contact mapped throughout the Ghanzi-Chobe zone. Similar methods could be used in other mineral districts with similar geology where cover obscures bedrock geology.

5.8 Conclusions

Stratigraphic analysis and analysis of numerically transformed high-resolution aeromagnetic maps from the Ghanzi Ridge area indicate that the intraformational aeromagnetic fabric of the 2nd VD map is controlled primarily by the contrasting grain size between lithostratigraphic units. The term magnetic lithostratigraphy is defined to describe the technique of identifying stratigraphic boundaries through the changes in the aeromagnetic map textures. The technique can be used as an indirect mapping
Figure 5.10 Interpretive cross-sections from sedimentary rock-hosted stratiform copper districts depicting the underlying basin architecture and the locations of orebodies. Note the difference in scale between the districts and orebodies. A) Kalahari Copperbelt, Botswana, Zone 5 and Northeast Mango Two orebodies. B) Kupferschiefer, Poland, Lubin-Sieroszowice and Kaleje orebodies. Modified from Hitzman et al. (2005). C) Central African Copperbelt (CACB), Zambia, Mwambashi B orebody. Modified from Selley et al. (2005). D) CACB, Zambia, Mufulira stacked orebodies. Modified from Hitzman et al. (2012).
tool, similar in sense to seismic stratigraphy, provided that the correct geological conditions are met (e.g. folded, moderately to steeply dipping and eroded (meta)sedimentary rocks and magnetically inert cover rocks). The 2nd VD map was used for along-strike correlation of magnetic lithostratigraphic boundaries/units for several kilometers on fold limbs. Additional structural information can be inferred through locating lineaments that correspond to minor offsets in, or destruction of, the 2nd VD aeromagnetic fabric. Resultant interpretations of the magnetic lithostratigraphy technique provided insights into the stratigraphic architecture of under- and overlying formations that have little to no available geologic data.

Magnetic lithostratigraphy analysis of the host rocks to the Zone 5 and Northeast Mango Two copper-silver deposits reveals several previously unrecognized geological features. The distinctive magnetic fabric of the Kuke Formation displays lateral thickness changes on onlap relationships that imply syn-sedimentary faults adjacent to paleotopographic highs. The overlying Ngwako Pan Formation is inferred to have filled a ~12-km-wide sub-basin between paleographic highs (Zone 5 and Makgabana Hills domes) based on the curvilinear concave up-down-up pattern observed in the 2nd VD aeromagnetic fabric. Above the inferred paleotopographic highs, the uppermost magnetic lithostratigraphic boundaries of the Ngwako Pan Formation are truncated along the contact with overlying D’Kar Formation. This indicates a previously unrecognized regional-scale unconformity that varies between an angular unconformity above paleotopographic highs where erosion occurred and a disconformity in the intervening basins where the erosional detritus was re-deposited and reworked during the marine transgression. Lateral facies changes that extend away from the paleotopographic highs indicate that the dome-like features persisted throughout the transgression. These paleotopographic highs are inferred to have been critical to the development of favorable litho-chemical traps sites and the fluid pathways for mineralizing fluids during subsequent basin inversion. The underlying basin architecture to the copper-silver deposits is similar to that documented in other world-class sedimentary rock-hosted copper districts around the world.

5.9 Acknowledgements

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5.10 References cited


CHAPTER 6
GENERAL CONCLUSIONS

The multi-disciplinary research of the Ghanzi basin and mineralizing systems of the Kalahari Copperbelt in Botswana provided insight into the two outstanding unknowns of the Kalahari Copperbelt. The multi-faceted geochronological investigations resolved the age of the host rocks to the Cu-Ag deposits as well as the provenance of the metasedimentary rocks of the Ghanzi Group. The geological and geophysical investigation aided in the identification of large-scale stratigraphic and basin architectural details that allowed for a reasonably well constrained interpretation of the tectonostratigraphic evolution of the Ghanzi basin. The location of Cu-Ag deposits was coincident with the locations of paleotopographic highs that were inferred from the stratigraphic and basin architectural details of the second vertical derivative (2nd VD) aeromagnetic map. The data suggests that the underlying basin architecture strongly influenced both the formation of physical and chemical traps as well as the fluid pathways used by mineralizing fluids. Re-Os sulfide and U-Th-Pb xenotime geochronology data indicate several pulses of mineralization in the Kalahari Copperbelt.

6.1 Resolving the age of the Ghanzi Group

LA-ICPMS U-Pb geochronology of igneous zircon (Chapter 3) provided a maximum depositional age constraint for the Ghanzi Group. The U-Pb zircon age of 1085.5 ± 4.5 Ma obtained from the Makgabana Hills rhyolite (see Section 3.6.1) is approximately 20-million-years younger than the 1106 ± 2 Ma Mabaleapodi Hills rhyolite, which is located ~70 kilometers to the southwest (ID-TIMS zircon; Schwartz et al., 1996). It provides a new maximum depositional age constrain for the unconformably overlying Ghanzi Group and the initiation of the northwest Botswana rift. Additionally, it expands the timing of volcanic activity in the northwest Botswana rift by ~20 Ma. Previously, U-Pb zircon ages constrained all of the bimodal volcanic rocks and related intrusions studied to this point in the northwest Botswana rift to the period between ~1112 to ~1100 Ma, which was associated with the Umkondo-Borg LIP event on the Kalahari Craton (Hegenberger and Burger, 1985; Schwartz et al., 1996; Singletary et al., 2003; Hanson et al., 2004; de Kock et al., 2014).

Application of LA-ICPMS to detrital zircon obtained from the sedimentary rocks of the Ghanzi Group (Chapter 3) established a maximum depositional age of the middle and upper Ghanzi Group (D’Kar and Mamuno formations, respectively). Weighted-mean ages of the youngest detrital zircon age populations from the uppermost Ngwako Pan (1066.4 ± 9.4 Ma), lower D’Kar (1063 ± 11 Ma), and Mamuno (1056.0 ± 9.9 Ma) formations constrained the timing of D’Kar Formation sedimentation to after ~1060 to ~1050 Ma (see Section 3.6.2).
To place a minimum depositional age constraint on the Ghanzi Group, and in particular the cupriferous D’Kar Formation, Re-Os and U-Th-Pb geochronology (Chapter 4) was employed on mineral phased that grew within (post-date deposition of) the host rocks. Previously reported Re-Os ages of ~995 to ~1017 Ma (bornite) and 914 ± 4 Ma (low-level highly radiogenic chalcopyrite) hinted at a possible depositional age for the D’Kar Formation, but the vein-hosted nature of the copper sulfides that were analyzed were difficult to reconcile with the available geological data. However, the textures observed during petrographic analysis of the molybdenite-bearing sample from the Northeast Mango Two deposit (see Section 4.4.2) provided strong evidence for the replacement of original diagenetic sulfide, framboidal pyrite, by hydrothermal molybdenite. Importantly, the age of the hydrothermal mineral assemblage was constrained to 538.5 ± 4.5 Ma by U-Th-Pb dating of a xenotime inclusion that was intergrown with molybdenite within vein-hosted chalcopyrite from the same sample. The combined petrographic and geochronological data suggest that the hydrothermal molybdenite replaced the framboidal pyrite in-situ and inherited, wholesale, the Re-Os systematics, and hence the age, of the framboidal pyrite grains. The newly reported molybdenite (after framboidal pyrite) Re-Os age of 981 ± 3 Ma established a minimum depositional age constraint for the lowermost, ore-hosting D’Kar Formation. The ~950 to 925 Ma U-Th-Pb ages obtained from several apparently bedded authigenic xenotime grains within the lowermost D’Kar Formation (Zone 6 deposit, see Section 4.6.3) support this.

The ~65 to ~75 million years between the maximum (~1060 to ~1050 Ma) and minimum age (981 Ma) of sedimentation of the D’Kar Formation overlaps the 1040-1020 Ma Klondikean episode of Namaquan orogenesis on the western margin of the Kalahari Craton, which reflects continent-continent collision with Laurentia during the assembly of Rodinia (Clifford et al., 2004; Miller, 2012; Swanson-Hysell et al., 2015). This contrasts to the timing of other Neoproterozoic sedimentary basins that host significant sedimentary rock-hosted copper deposits, which initiated during the break-up of Rodinia after ~900 Ma.

6.2 Provenance of the Ghanzi Group

Lu-Hf isotopic analysis to the detrital zircon suite (see Section 3.6.3) helped characterize the nature of those zircon groups and sub-groups within the detrital zircon U-Pb geochronology data. The new data combined with a compilation of magmatic U-Pb zircon ages along with Lu-Hf and Sm-Nd model ages (see Section 3.7.2) from the literature on the Precambrian terranes of the Kalahari Craton suggested that the western margin of the Kalahari Craton, comprised of the Paleo- to Mesoproterozoic Namaqua Sector and the Rehoboth Basement Inlier, was the primary sources for the Ghanzi Group metasedimentary rocks. The Paleoproterozoic rocks that bound the southern margin of the Ghanzi basin (northwest Botswana rift) were a minor sediment source for the Ghanzi Group.
6.3 Stratigraphic and basin architecture of the Ghanzi basin

High-resolution aeromagnetic surveys carried out by mineral exploration companies offered an opportunity to investigate the aeromagnetic signature of the Ghanzi Group. Numerical transformations of the grid data revealed the intraformational magnetic fabric of the Ghanzi Group rock formations (see Section 5.4.2.1). The 2nd VD transformation resolved the near surface, high frequency signal of the underlying rock formations. Comparison of well-constrained D’Kar Formation lithostratigraphic sections to the 2nd VD aeromagnetic maps revealed a strong correlation between coarser-grained (sandstone) lithostratigraphic units and the strike-parallel, elongate textured, high magnetic intensity anomalies; finer-grained (mudstone-siltstone-carbonate) lithostratigraphic units corresponded to low intensity anomalies (see Sections 5.4.1.2 and 5.4.2.2). Recognition of this relationship between lithostratigraphic units and their corresponding aeromagnetic fabric and the resultant geological interpretations is proven as a new remote mapping technique termed magnetic lithostratigraphy, which is interpreted in a similar fashion to seismic stratigraphy (see Section 5.5).

Magnetic lithostratigraphy was used to extrapolate the stratigraphic architecture of the Ghanzi Group on steeply dipping fold limbs for several kilometers along strike (see Section 5.5). The resultant magnetic lithostratigraphy interpretation revealed several previously unrecognized stratigraphic, structural, and basin architectural details. Onlap surfaces and lateral thickness changes within the basal Kuke Formation and overlying Ngwako Pan Formation suggest the presence of syn-sedimentary faults adjacent to paleotopographic highs. The concave up-down-up pattern formed by the magnetic lithostratigraphic units of the Ngwako Pan Formation suggest that paired (relative) uplift and subsidence affected the Ngwako Pan Formation during deposition. Truncation of the magnetic lithostratigraphic units of the Ngwako Pan Formation above and adjacent to the inferred basement highs suggests that erosion of these features occurred prior to marine incursion. Lateral facies shifts within the base of the overlying D’Kar Formation confirm that the paleotopographic highs persisted during marine incursion. The result was the formation of favorable physical and chemical trap sites above inferred syn-sedimentary, sub-basin bounding faults. The faults that are suggested by the magnetic lithostratigraphy seem to be responsible for channeling several pulses of mineralizing fluids into the base of the D’Kar Formation during basin inversion to form the structurally-controlled sedimentary rock-hosted copper-silver deposits of the Kalahari Copperbelt.

In addition to magnetic lithostratigraphy, the 2nd VD map revealed at least two orientations of major structures that cross-cut the folded Ghanzi Group. The structural trends may also help in the interpretation of the location and attitude of inferred syn-sedimentary faults given enough geological information. The extracted structural trends could also aid in structural modelling at the deposit scale.
6.4 Mineralizing systems in the Kalahari Copperbelt

The Re-Os and U-Th-Pb geochronology of sulfides and phosphate minerals, respectively, indicate two periods of mineralization in the Kalahari Copperbelt: diagenetic-aged event(s) spanning the time period of ~1000 to 900 Ma and epigenetic-aged events that span the time interval from ~680 to ~450 Ma (see Section 4.7).

Combined petrographic and geochronological evidence suggest that hydrothermal molybdenite replaced bedded concentrations of diagenetic framboidal pyrite grains during hydrothermal fluid-assisted shearing of wall rock fragments. The syngenetic to diagenetic-aged Re-Os sulfide ages that have been obtained from the D’Kar Formation occur in close proximity to presumed paleotopographic highs. The Northeast Mango Two Cu-Ag deposit (981 ± 3 Ma molybdenite) occurs adjacent to the Makgabana Hills paleotopographic high. The Zeta Cu-Ag deposit (995 to 1027 Ma bornite and 914 ± 4, LLHR chalcopyrite, see Chapter 4.5-4.6 for explanation; Hall, 2013) occurs adjacent to the Kgwebe Hills paleotopographic high. These paleotopographic highs may have influenced the intrabasinal depositional environments and localized anoxic sub-basins that were favorable to bacteriogenic sulfate reduction.

The ~670 to ~630 Ma arsenopyrite Re-Os age range obtained from the Zone 5 deposit represents a previously undocumented mineralizing event within the Ghanzi basin and the northwest Botswana rift. The range coincides with the end of the global “Snowball Earth” Sturtian glaciogenic event and the later ~635 Ma Marinoan glaciogenic event (e.g. Hoffman et al, 1998, Miller, 2008 and references therein). The age range is difficult to reconcile with the currently available geological and geochronological data for the Kalahari Copperbelt, but the coarse-grained prismatic texture of the grains is akin to diagenetic coarse-grained cubic pyrite that occurs within the rocks as well. However, the petrographic evidence indicates that the coarse-grained arsenopyrite mineralizing event preceded the main Cu-bearing pulses of epigenetic, syn-kinematic mineralization in the Kalahari Copperbelt.

The combined Re-Os sulfide and U-Th-Pb xenotime ages (549.0 ± 11.2 Ma, 538.4 ± 8.3 Ma, 515.9 ± 2.4 Ma, and 442 to 496 Ma) obtained from mineralized veins in the Kalahari Copperbelt suggests that several pulses of epigenetic base metal mineralization occurred within the Ghanzi-Chobe zone during the prolonged episode of Damara orogenesis. Multiple pulses of mineralizing events in the Kalahari Copperbelt are reflected in the abundant cross-cutting relationships between mineralized structures that characterize the structurally controlled ore deposits.

6.5 General conclusions

The research has important implications for the field of Re-Os geochronology in that Re-Os systematics of sulfides can survive in-situ on solid state recrystallization to other sulfide species. Detailed petrographic and geochronologic work should be employed in any situation where solid state replacement of one sulfide species by another has taken place.
Magnetic lithostratigraphy is defined to describe the technique of identifying stratigraphic boundaries through the changes in the aeromagnetic map textures. The technique can be used to map stratigraphy in covered, deformed sedimentary basins provided that the correct geological conditions are met (e.g. folded, moderately to steeply dipping and eroded (meta)sedimentary rocks and magnetically inert cover rocks). Ground truthing of the technique should be carried out from well-constrained areas prior to wide-spread application of the technique. Magnetic lithostratigraphy can be interpreted in similar fashion to seismic stratigraphy and the same cautions should be employed when using the technique.

6.6 References cited


APPENDIX A

ADDITIONAL PETROGRAPHY AND INCONCLUSIVE RE-OS RESULTS

This appendix contains additional petrographic work as well as inconclusive Re-Os geochronology results from two samples. The first sample was obtained from the Makgabana Hills bimodal volcaniclastic rocks (see Chapter 3 for location) and consisted of vein-hosted chalcopyrite with inconclusive Re-Os results. The second sample consisted of a second separate from the Zone 6 quartz-calcite-chalcocite-idaite-bornite-native silver vein which contained common Os and thus an inconclusive age that cannot be reconciling within the framework of this project. Petrography was also carried out on a third sample from the Zeta Cu-Ag deposit that yielded a Re-Os age from a previous study.

A-1 Makgabana Hills (The Dome), chalcopyrite (AIRIE Run # MD-1362)

The Makgabana Hills (the Dome prospect) contains veinlet and disseminated chalcopyrite within the Kgwebe Formation. The host rocks consist of dark gray porphyritic rhyolite flows with a distinct discontinuous flow foliation. The rhyolites are intercalated with volcaniclastic deposits that contain rounded sedimentary clasts and wispy to layered injections of metasediments (Figure A-1). The rhyolite is composed of 1-3 mm diameter feldspar phenocryst set in an aphanitic quartz-feldspar groundmass. The rhyolites are commonly cut by thin, dark, magnetite-bearing fractures and veinlets with thin to thick, reddish colored hematitic alteration selvages (Figure A-1). These fractures appear to be coeval with quartz-(carbonate-magnetite-chalcopyrite) veins based on a lack of cross-cutting relationships and the presence of weak to strong hematitic alteration selvages to both veinlets and veins (Figure A-1). Hematitic alteration within both vein sets is more apparent in volcaniclastic lithologies, probably due to their higher inherent permeability (Figure A-1).

A 2-cm-thick chalcopyrite-magnetite vein hosted within a dark red auto-brecciated rhyolite from drill hole DMDD2183 at a depth of 68.8 meters was selected for Re-Os isotopic analysis (Figure A-1). The chalcopyrite is intergrown with and encloses magnetite within the quartz vein. Petrography indicates that the phenocrysts are primarily albite that have variably altered to potassium feldspar (Figure A-2). The groundmass is composed primarily of albite and potassium feldspar and quartz; EDS analysis indicated that the potassium feldspar consistently contains minor Na, which suggests potassic alteration of an originally albitic groundmass. The rhyolite commonly contains aphanitic albite-quartz and clots of manganese-rich epidote and chlorite that texturally resemble amygdules. The albite-epidote-chlorite (Na-Ca) assemblage appears to be strongly overprinted by a potassic alteration phase that resulted in potassium feldspar flooding of the groundmass and replacement of albrite and epidote by potassium feldspar and biotite (Figure A-3); these potassium-bearing minerals are commonly intergrown with magnetite-hematite, titanite, monazite, and minor LREE-bearing minerals (Figure A-3). EDS analysis
Figure A-1 Rhyolite host rocks to basement-hosted copper mineralization, the Dome. A) Largely unaltered gray porphyritic rhyolite flow with minor dark magnetite-bearing veinlets with red hematitic alteration selvages (arrow). B) Volcanoclastic unit with abundant wispy to bedded injections of metasediments. The unit displays strongly hematitic alteration. C) Moderately altered rhyolite cut by quartz-carbonate veins and chalcopyrite-magnetite vein (arrow) sampled for Re-Os isotopic analysis. D) Macro photo of the chalcopyrite-magnetite vein, hosted in a rhyolite with a breccia texture.
Figure A-2 BSE image displaying a magnetite-bearing fracture in porphyritic rhyolite consisting of Na-plagioclase phenocrysts set in an albite-K-feldspar groundmass. Larger albite feldspar phenocrysts adjacent to the magnetite-bearing fracture have been replaced by epidote followed by magnetite-ilmenite-K feldspar. K-feldspar and magnetite alteration is ubiquitous around the fracture. Epidote is concentrated along discontinuous flow foliations that are cut by the magnetite-bearing fracture.

indicates that the potassium feldspar commonly contains minor amounts of barium. Potassium feldspar flooding of the groundmass is associated with thin dark fractures/veinlets containing abundant magnetite and minor chalcopyrite and bornite that display reddish alteration selvages containing finely disseminated hematite (Figures A-2 and A-3).

Chalcopyrite from the Dome prospect (MD-1362) was separated utilizing a hand-held drill and yielded 95% pure chalcopyrite mineral separate weighing 302 milligrams. The sample was combined with a double-Os spike for mass fractionation and determination of common Os. Re and Os concentrations in the chalcopyrite are exceedingly low (less than 2 ppb Re and less than 2 ppt Os; Table A-1). MD-1362 was overspiked, but with Re levels in the vicinity of 2 ppb, the data were highly
Figure A-3. BSE images showing alteration phases in porphyritic rhyolite. A) A large subhedral phenocryst of albite (yellow outline, darker gray) that is weakly altered to K-feldspar (lighter gray). Inset location refers to image (B). B) Alteration within an albite phenocryst. Albite is replaced by K-feldspar and epidote + magnetite ± titanite. C) Albite feldspar with epidote that is replaced by K-feldspar and biotite. D) Intergrown epidote, magnetite, and a Ca-Fe-Br-Ce-Nd aluminosilicate mineral surrounded by K-feldspar that replaces albite. E) Feldspar grains that have been replaced by a magnetite-ilmenite-epidote-monazite assemblage. F) Close-up of (E) displaying exsolution texture between magnetite and ilmenite that is intergrown with epidote, titanite, and monazite.
Table A-1 Inconclusive or non-determinable Re-Os results from this study

<table>
<thead>
<tr>
<th>AIRIE Run #</th>
<th>Sample ID (Drill Hole ID, depth, prospect, sulfide)</th>
<th>Re, ppm</th>
<th>Re error, abs (ppm)</th>
<th>$^{187}$Os, ppb</th>
<th>$^{187}$Os error, abs (ppb)</th>
<th>Common Os, ppb</th>
<th>Common Os error, abs (ppb)</th>
<th>Age, Ma</th>
<th>Age Error, Abs (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD-1362</td>
<td>DMDD-2183, 68.8 m, “The Dome” chalcopyrite</td>
<td>0.002</td>
<td>0.001</td>
<td>0.001639</td>
<td>0.000003</td>
<td>0.00069</td>
<td>0.00005</td>
<td>ND</td>
<td>ND</td>
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<tr>
<td>MD-1696</td>
<td>HA-641-D, 501.1 m, “Zone 6” Cc-Id-Bn-native Ag (new sep.)</td>
<td>0.01587</td>
<td>0.0000017</td>
<td>0.036973</td>
<td>0.000054</td>
<td>0.06967</td>
<td>0.000014</td>
<td>222.1</td>
<td>0.8</td>
</tr>
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</table>

Cc = chalcocite, Id = idaite, Bn = bornite, Qtz = quartz, Cal = calcite, ND = not determinable, new sep. = new mineral separate
All uncertainties reported as two-sigma, uncertainty in Re-Os ages include the $^{187}\text{Re}$ decay constant uncertainty.
All sample equilibrated with spike using Carius tube method; Re and Os isotopic ratios measured by NTIMS on a Triton machine at the AIRIE Program.
Sample weights are 302 mg for chalcopyrite (MD-1362) and 20.153 mg for the Cc-Id-Bn-native Ag mixture (MD-1696).
For MD-1362, separate is 95% chalcopyrite (i.e. 5% silicate dilution). Silicate dilution does not affect Re-Os ages.
MD-1362 is overspiked, but with Re levels in the vicinity of 2 ppb, data will be highly sensitive to the blank correction and therefore it will be difficult to derive a good isochron.
Assumed initial $^{187}$Os/$^{188}$Os for age calculation for MD-1696 = 0.2; assumed Os initial includes 1-% uncertainty.
For MD-1696, mineral separate included a ductile mineral that was malleable and coated the diamond drill bit, likely native silver, 40% metallic component in separate; mixture of events, age is almost certainly meaningless.
For MD-1362, Re blank = 3.814 ± 0.39 pg, Os blank = 0.174 ± 0.012 pg with $^{187}$Os/$^{188}$Os = 0.313 ± 0.017.
For MD-1696, Re blank = 5.510 ± 0.01 pg, Os blank = 0.791 ± 0.008 pg with $^{187}$Os/$^{188}$Os = 0.420 ± 0.004.
sensitive to the blank correction; therefore, the Os is so low that the sample is not datable by the Re-Os method (Table A-1).

A-2 Second separate from Zone 6, Qtz-Cal-Cc-Id-Bn-native Ag vein (AIRIE Run# MD-1696)

Because the first chalcocite-idaite separate had a molybdenite-like Os isotopic composition (see Section 4.6.1), a second mineral separate was obtained and ran utilizing a double Os spike. The sample contained common Os (MD-1696; Table A-1). Native silver was identified in the new sample separate, which could contain common Os. When common Os is present, the Re-Os age is dependent on the choice the initial Os ratio. For MD-1696, any realistic Os initial ratio results in a range of ages between 225 Ma (chondritic initial Os = 0.12) and 125 Ma ($^{187}$Os/$^{188}$Os = 2). The data suggests that the vein mineral assemblage Re-Os systematics may have been affected by a Jurassic-aged event and therefore cannot be reconciled within the context of this study. The age range overlaps with the Karoo-aged dolerite dykes of the Okavango dyke swarm that cross-cut the Ghanzi-Chobe zone and are located within a few kilometers of the Zone 6 prospect.

A-3 Zeta deposit, GDRD1127_262 m, vein-hosted chalcopyrite (AIRIE Run # LL-611)

Petrography was conducted on a sample of vein-hosted chalcopyrite that was dated at 914 ± 4 Ma (Hall, 2013). The petrography aimed to identify molybdenite and other trace minerals that were observed in the current study (Bi, Ni, Co, As, LREE elements). Molybdenite was not observed as a primary constituent of the quartz-calcite-chalcopyrite vein. However, abundant micron-sized molybdenite grains were observed within the adjacent phengitic wall rock (Figure A-4). It forms small trails parallel to lithological layering within the wall rock (Figure A-4). Minor molybdenite occurred at the vein-wall rock boundary where it was intergrown with chalcopyrite-(bornite) and Ca-bearing LREE fluoro-carbonate.
minerals, similar to the mineralogical relationships observed at other deposits. Molybdenite is also intergrown with rutile in the sample (Figure A-4).

A-4 References cited

APPENDIX B
AEROMAGNETIC SURVEY SPECIFICATIONS

The following data table outlines the specifications of the aeromagnetic survey that were used for the qualitative comparison of lithostratigraphy to aeromagnetic fabrics in Section 5.3. The district-scale high-resolution helicopter-borne aeromagnetic surveys were transformed and utilized for the magnetic lithostratigraphy interpretation.

Table B-1 Flight data acquisition parameters and specifications for individual aeromagnetic surveys within the Ghanzi-Chobe zone of northwestern Botswana (see Section 5.3)

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<th>Survey Name</th>
<th>Line spacing</th>
<th>Tie line spacing</th>
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<th>Average flight height (m)</th>
<th>Year of Survey</th>
<th>Collected by</th>
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<td>N/A</td>
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<td>District-Scale Helicopter-Borne Aeromagnetic Surveys</td>
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<td>NNW-SSE</td>
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Petrophysical samples were collected from the Zone 5 copper-silver deposit in the Northeast District of the Kalahari Copperbelt in Botswana. Quantitative analysis of the different lithostratigraphic units of the D’Kar Formation was undertaken to numerically model the induced intensity of magnetization of the rocks, the results of which were compared to the qualitative analysis between well-constrained lithostratigraphy and second vertical derivative aeromagnetic maps (Chapter 5). Measurements of anisotropy of magnetic susceptibility (AMS) and natural remanent magnetization (NRM) were conducted by Dr. Zheng Zhong of the Geoscience Division of Sogokaihatsu Company Ltd., Japan.

C-1 Petrophysical properties

Magnetic susceptibility is a dimensionless parameter that quantifies the degree of magnetization of a material in response to an applied magnetic field. Magnetic susceptibility is defined as $\kappa = M / H$ where $\kappa$ is the magnetic susceptibility, $M$ is the induced magnetization per unit volume, and $H$ is the applied magnetic field (Tarling and Hrouda, 1993). The magnitude of $\kappa$ in polymineralic rocks depends on the nature and concentration of diamagnetic, paramagnetic, and ferromagnetic minerals present. The three orthogonal components (magnitude and vector orientation) of $\mathbf{J}_i$ can be measured to determine if there is a preferred orientation and intensity of magnetic grains in rocks. This is termed anisotropy of magnetic susceptibility (AMS) and can be used to define the magnetic fabric of rock (see Tarling and Hrouda, 1993 for review; Elmore et al., 2016). AMS is represented by a symmetric second rank tensor with three mutually perpendicular principal axes $\kappa_{\text{max}}$ (long axis), $\kappa_{\text{int}}$ (intermediate axis), and $\kappa_{\text{min}}$ (short axis). In addition, the corrected degree of anisotropy ($P' = \kappa_{\text{max}} / \kappa_{\text{min}}$; Jelinèk, 1981), a proxy for mineral shape anisotropy (i.e. the degree to which the AMS ellipsoid deviates from a sphere), and shape factor ($T$; Jelinèk, 1981), that describes the shape of the AMS ellipsoid with end-members ranging from oblate ($0 \leq T \leq 1$), to spherical ($T = 0$) to prolate (-1 $\leq T \leq 0$), can be computed (see Jelinek, 1981; Heij et al., 2016). AMS fabrics generally show good correlations with certain mineralogical end members. Planar fabrics are due to magnetostatic (shape) anisotropy due to mineral such as, but not limited to, clays, chlorite, biotite, and pyrite. High inclination fabrics may occur as the carbonate fraction increases due to the magnetocrystalline anisotropy inherent to diamagnetic ferroan carbonate minerals. As the silica fraction increases, an exponential increase in the fabric inclination occurs while near vertical fabrics are commonly associated with slumped beds and soft sediment deformation (Heij et al., 2016).

Anisotropy of magnetic susceptibility (AMS) was measured to investigate the relationships between lithology, deformation, and hydrothermal alteration and the resulting magnetic petrofabrics (i.e.
planar versus vertical magnetic fabrics and orientation in relation to cleavage and/or foliation; e.g. Girdler, 1961; Uyeda et al., 1963; Rees, 1965; Jelinek, 1977, 1981; Borradaile, 1987, 1991; Borradaile and Henry, 1997; Borradaile and Jackson, 2004) in several metasedimentary units within the lower D’Kar Formation. Measurements of the natural remanent magnetization (NRM) were carried out to determine if any NRM was present within the rocks and if so, its orientation (e.g. Girdler and Peter, 1960). AMS and NRM measurements were conducted on six samples of mineralized and non-mineralized, fine- and coarse-grained rock types from stacked lithological units of the D’Kar Formation at the Zone 5 deposit. The samples consisted of six-cm-long intervals of oriented NQ (47.6 mm diameter) diamond drill core. AMS measurements were conducted using an AGICO KLY3 Kappabridge susceptibility meter and standard NRM measurements were carried out on a AGICO JR-5A magnetometer by the Geoscience Division of Sogokaihatsu Company Ltd, in Tokyo, Japan. Scalar parameters were obtained from the AMS measurements using SUSAR software from AGICO. AMS and NRM directional data were plotted on equal area stereonets using the program GEOrient© (Rod Holcombe). The parameters used are mean susceptibility (average of the three principal axes of the susceptibility ellipsoid: $\kappa_m = \kappa_{\text{max}} + \kappa_{\text{int}} + \kappa_{\text{min}} / 3$), the magnetic foliation parameter (relation between the intermediate and minimum susceptibility axes; $F = \kappa_{\text{int}} / \kappa_{\text{min}}$), the magnetic lineation parameter (relation between the long an intermediate susceptibility axis: $L = \kappa_{\text{max}} / \kappa_{\text{min}}$), the shape parameter (T) and corrected degree of anisotropy, $P'$, following the definition of Jelinek (1981).

**C-2 Petrophysical results**

To better understand the relationships between lithostratigraphy and the 2nd VD fabrics, the intensity of magnetization was computed and modelled for the lowermost lithological units of the D’Kar Formation at Zone 5 (Table C-1). This was accomplished through measuring the magnitude of the induced magnetization (or magnetic susceptibility, $J_i$) and vector magnitude of the permanent natural magnetic remanence (NRM, $J_r$, due to the internal field strength of permanently magnetic particles). After multiplying the $J_i$ by the Earth’s magnetic field, the two vector components $J_r$ and $J_i$ were summed to determine the resultant induced intensity of magnetization ($J$), which dictates the both the amplitude and shape of a magnetic anomaly.

In order to determine if a specific magnetic fabric was imparted to the rocks, the resultant $\kappa_{\text{max}}$, $\kappa_{\text{int}}$, and $\kappa_{\text{min}}$ AMS principal directions for each sample were rotated to the true downhole orientation of the sample and plotted on equal-area lower hemisphere projections; Figure C-1). Bulk magnetic susceptibilities of the lower D’Kar Formation rocks have a narrow range from $\sim 3.0 \times 10^{-3}$ to $\sim 5 \times 10^{-4}$ SI units (see Section 5.4.2.1). Therefore, the distinct magnetic fabrics within the fine- (and often carbonate-rich) and coarse-grained lithological units could give rise to distinctive variations in the resultant intensity of
Table C-1 Petrophysical sample locations and descriptions

<table>
<thead>
<tr>
<th>Drill Hole ID (Zone5)</th>
<th>Depth</th>
<th>Azimuth/ Dip</th>
<th>Stratigraphic position</th>
<th>Lithology</th>
<th>Bedding (S₀) /Foliation (S₁) (RHR Strike/Dip)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>HA-1056-D</td>
<td>206.55 – 206.61</td>
<td>321.3/- 58.0</td>
<td>Sandstone Unit A</td>
<td>Sandstone</td>
<td>S₀: N/A S¹: N/A</td>
<td>Grey, massive bedded, fine- to very fine-grained sandstone/greywacke cut by thin quartz-calcite ± sphalerite veinlets and cross-cutting calcite-hematite veinlets, very weakly magnetic</td>
</tr>
<tr>
<td>HA-1056-D</td>
<td>225.55 – 225.61</td>
<td>321.1/- 56.9</td>
<td>Alternating Unit A</td>
<td>Siltstone</td>
<td>S₀: N/A S¹: 056/73</td>
<td>Black, thinly bedded graphitic siltstone with finely disseminated pyrite cut by quartz-carbonate veins overprinted by a cataclasite shear structure, non-magnetic</td>
</tr>
<tr>
<td>HA-1056-D</td>
<td>240.55 – 240.61</td>
<td>320.5/- 56.5</td>
<td>Alternating Unit A</td>
<td>Siltstone</td>
<td>S₀: 064/52 S¹: 060/73</td>
<td>Grey to dark grey, thinly bedded, normally graded to ripple laminated siltstone containing blebby-nodular pyrite along bedding contacts and disseminated in sandier laminations, non-magnetic</td>
</tr>
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<td>320.7/- 55.6</td>
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<td>Sandstone</td>
<td>S₀: N/A S¹: 055/70</td>
<td>Grey, thin bedded, fine- to very fine-grained sandstone/greywacke</td>
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<td>Upper Ore Zone Unit</td>
<td>Marlstone</td>
<td>S₀: 050/53 S¹: 046/83</td>
<td>Light blue-grey, thin bedded to laminated marlstone cut by thin, dark foliation planes containing blebby pyrite-chalcopyrite ± quartz-calcite, non-magnetic</td>
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<td>Siltstone</td>
<td>S₀: 057/56 S¹: 051/69</td>
<td>Grey to dark grey, thin bedded and internally laminated siltstone with blebby disseminated pyrite concentrated within coarser laminations, non-magnetic</td>
</tr>
</tbody>
</table>

magnetization ($J$), which can be modeled in profiles and compared to the observed aeromagnetic response of the D’Kar Formation rocks.

C-2.1 Anisotropy of magnetic susceptibility (AMS)

Measurement of the normed principle components of magnetic susceptibility were used to determine the degree of anisotropy ($P'$) and the shape factor (T) for all samples (Table C-2). The $P'$ calculated for all samples varies from 1.011 to 1.123, indicating that a weak (~1.0) to strong (>1.0) anisotropic fabric was imparted to all of the rocks (Table C-2). Sandstone units have a more spherical shape (T = 0.037 to 0.145) while fine-grained rocks having more oblate shapes (T= 0.327 to 0.810; Table C-2).

The $\kappa_{max}$-$\kappa_{int}$ planes for the majority of the samples define a magnetic foliation plane that strikes parallel to that of the mean bedding and mean foliation (approximately ~055°) measured throughout the Zone 5 deposit area (Figure C-2). The magnetic foliation plane dips ~60° to the northwest, opposite that of bedding and foliation, which dip ~60° and ~75-80° to the southeast, respectively (Figure C-2). The exceptions to this are the undeformed Sandstone Unit A (HA-1056-D_206.6 m), which has an apparent magnetic foliation that strikes slightly more to the north (strike of ~035 to 040° as opposed to 230 to 240° for bedding and foliation; Figure C-2), and one set of measurements from the Alternating Unit A sheared carbon-rich siltstone (HA-1056-D 225.1 m), which has an east-west striking, near vertical (89°) fabric.
Figure C-1 AMS principal susceptibilities and NRM vector data, rotated to true orientation.  A) Core photographs of petrophysical samples displaying bedding ($S_0$) and foliation ($S_1$).  B) Equal-area upper (top) and lower (bottom) hemisphere projections (top and bottom, respectively) of NRM vectors (triangles), AMS principle directions (colored circles), planes representing $\kappa_{\text{max}}$-$\kappa_{\text{int}}$ (red) plane that defines the AMS ellipsoid, $S_0$ (black), and $S_1$ (yellow).
Table C-2 Anisotropy of magnetic susceptibility results (true orientations)

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Core Coordinates</th>
<th>Normed Principal Susceptibilities</th>
<th>Degree of magnetic anisotropy</th>
<th>Shape Factor</th>
<th>Magnetic Susceptibility (x10^-6 SI units)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\kappa_{\text{max}}$</td>
<td>$\kappa_{\text{int}}$</td>
<td>$\kappa_{\text{min}}$</td>
<td>$\kappa_1$</td>
<td>$\kappa_2$</td>
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<tr>
<td>206.63 - 207.00 Sandstone</td>
<td>220 -13 16 -77</td>
<td>310 -32</td>
<td>1.0343</td>
<td>0.9983</td>
<td>0.9674</td>
</tr>
<tr>
<td>(Sandstone Unit A)</td>
<td>218 -15 354 -70</td>
<td>311 04</td>
<td>1.0318</td>
<td>0.9989</td>
<td>0.9693</td>
</tr>
<tr>
<td>225.43 - 225.50 Veined, sheared</td>
<td>172 -66 61 -10</td>
<td>321 -23</td>
<td>1.0129</td>
<td>1.0095</td>
<td>0.9775</td>
</tr>
<tr>
<td>siltstone (Alternating Unit A)</td>
<td>068 -35 204 -58</td>
<td>321 -12</td>
<td>1.0050</td>
<td>1.0014</td>
<td>0.9936</td>
</tr>
<tr>
<td></td>
<td>086 -49 270 -43</td>
<td>359 04</td>
<td>1.0083</td>
<td>0.9983</td>
<td>0.9933</td>
</tr>
<tr>
<td>240.50 - 240.56 Siltstone</td>
<td>130 -60 232 -12</td>
<td>328 -27</td>
<td>1.0424</td>
<td>1.0292</td>
<td>0.9284</td>
</tr>
<tr>
<td>(Alternating Unit A)</td>
<td>134 -70 236 -04</td>
<td>328 -20</td>
<td>1.0355</td>
<td>1.0164</td>
<td>0.9481</td>
</tr>
<tr>
<td>276.65 - 276.71 Sandstone</td>
<td>52 -6 159 -67</td>
<td>319 -21</td>
<td>1.0341</td>
<td>1.0015</td>
<td>0.9645</td>
</tr>
<tr>
<td>(Marker Sandstone)</td>
<td>51 -9 189 -59</td>
<td>319 -04</td>
<td>1.0302</td>
<td>0.9970</td>
<td>0.9728</td>
</tr>
<tr>
<td>341.17 - 341.23 Marlstone</td>
<td>121 -64 239 -8</td>
<td>322 -25</td>
<td>1.0185</td>
<td>1.0089</td>
<td>0.9726</td>
</tr>
<tr>
<td>(Ore Zone Unit)</td>
<td>114 -65 230 -10</td>
<td>324 -22</td>
<td>1.0229</td>
<td>1.0106</td>
<td>0.9665</td>
</tr>
<tr>
<td>342.30 - 342.36 Siltstone</td>
<td>196 -51 65 -27</td>
<td>321 -24</td>
<td>1.0223</td>
<td>1.0144</td>
<td>0.9633</td>
</tr>
<tr>
<td>(Ore Zone Unit)</td>
<td>197 -51 67 -29</td>
<td>322 -24</td>
<td>1.0224</td>
<td>1.0130</td>
<td>0.9646</td>
</tr>
</tbody>
</table>

Figure C-2 Equal-area lower hemisphere projections of bedding ($S_0$), foliation ($S_1$) and magnetic foliation defined by the alignment of $\kappa_{\text{max}}$-$\kappa_{\text{int}}$ planes for all samples (colored symbols = poles to planes). The magnetic foliation dips in the opposite direction of bedding and forms an acute $\sim 25^\circ$ angle with the tectonic foliation and wider $\sim 50$-60$^\circ$ angle with bedding. Note that the $\kappa_{\text{max}}$-$\kappa_{\text{int}}$ plane for the Sandstone Unit A massive sandstone sample does not align with the majority of the samples, suggesting that a magnetic foliation was not developed in this sample.
Interestingly, the $\kappa_{\text{max}}$ for all the fine-grained rocks describe a negative (out of the Earth) magnetic lineation within the foliation plane that is oriented parallel to overall tectonic shortening direction (NW-SE).

Several of the petrophysical samples contained a planar, weakly to strongly oblate shaped AMS fabric, with respect to bedding. The lithological control on the AMS fabric is indicated by the higher degrees of anisotropy ($P'$) and more oblate shapes factors (T) for the fine-grained (chlorite-biotite-muscovite-illite-bearing) rocks compared to the low $P'$ values and more spherical shapes contained within the coarser-grained rocks. In samples with a measured foliation, the $\kappa_{\text{max}}$-$\kappa_{\text{int}}$ planes define a magnetic foliation. A magnetic lineation (clustering of $\kappa_{\text{max}}$ principle susceptibilities) was determined to be present in the foliated fine-grained rocks; the lineation is stretched parallel to the overall tectonic shortening direction.

### C-2.2 Natural remanent magnetization (NRM)

Most of the Zone 5 petrophysical samples have negative NRM inclinations (upward out of the earth; Table C-3; solid data points in Figure C-1) and NRM declinations within the northwestern quadrant of the equal area stereonet (upper hemisphere projections; Figure C-1) with magnitudes that vary from \(\sim 3.0 \times 10^{-5}\) to \(\sim 7 \times 10^{-4}\) amperes/meter (A/m; Table C-2). The siltstone and adjacent marlstone from the Ore Zone Unit both have NRM orientations that are directed towards the north to northwest at negative inclinations between \(-15^\circ\) and \(-30^\circ\) and magnitudes between \(9.46 \times 10^{-5}\) A/m (siltstone) and \(6.39 \times 10^{-4}\) A/m (marlstone). The NRM for the overlying Marker Sandstone Unit has a westerly declination with negative inclinations of \(-45^\circ\) to \(-57^\circ\) and magnitude varying from \(1.41 \times 10^{-4}\) to \(4.26 \times 10^{-4}\) A/m. The siltstone from Alternating Unit A has an NRM with magnitudes between \(7.09 \times 10^{-5}\) and \(5.18 \times 10^{-4}\) A/m that varies from being oriented nearly flat and directed towards the north to a northwesterly declination with a steep negative inclination of nearly \(-70^\circ\). Another siltstone from the same unit that was strongly affected by hydrothermal fluid flow (quartz-calcite-pyrite veins) and brittle-ductile shear deformation has inclinations of \(-45^\circ\) to \(-57^\circ\) and magnitude varying from \(1.41 \times 10^{-4}\) to \(4.26 \times 10^{-4}\) A/m. The siltstone from Alternating Unit A has an NRM with magnitudes between \(7.09 \times 10^{-5}\) and \(5.18 \times 10^{-4}\) A/m that varies from being oriented nearly flat and directed towards the north to a northwesterly declination with a steep negative inclination of nearly \(-70^\circ\). Another siltstone from the same unit that was strongly affected by hydrothermal fluid flow (quartz-calcite-pyrite veins) and brittle-ductile shear deformation has drastically different orientations from other samples. The NRMs obtained for this rock appear to lie within a near vertical plane that strikes parallel to bedding, similar to the orientation of the steep tectonic foliation imparted the folded rock (Figures C-1 and C-2). The NRMs in this case have magnitudes of \(2.91 \times 10^{-5}\) and \(1.04 \times 10^{-4}\) A/m and positive (into the Earth) inclinations of \(55^\circ\) to \(70^\circ\), in stark contrast to the negative inclinations of the rest of the rock samples. The NRM for the overlying Sandstone Unit A
has a northwesterly declination, negative inclinations of -30° to -45°, and magnitudes of $1.9 \times 10^{-4}$ to $2.5 \times 10^{-4}$ A/m.

C-2.3 The total magnetization ($J$)

The AMS with three principal susceptibilities ($\kappa_{\text{max}}$, $\kappa_{\text{int}}$, and $\kappa_{\text{min}}$) leads to an induced magnetization whose direction deviates from the inducing field for the magnetization and is the product of the susceptibility tensor and the inducing magnetic field vector. The measured susceptibility values and the known earth’s magnetic field allow calculation of this part of the magnetization. The sum of the calculated induced magnetization and the measured NRM data for each sample yields the total magnetization in the unit. At this time, the total magnetization calculations have not been finalized.

C-3 References cited


The supplemental electronic file contains additional images and data tables from Chapter 3, organized by tabs. The file contains images of the sedimentary rock samples and one igneous sample utilized for U-Pb geochronology. The data tables include the LA-ICPMS U-Pb and Lu-Hf results from the detrital zircon study, as well as the corresponding Terra-Wasserburg concordia diagrams for the detrital zircon results. Annotated cathodoluminescence (CL) images for all four detrital zircon samples and the one igneous zircon samples are included with the file. The data table and reference list (separate tabs) utilized for the provenance study are also included.

<table>
<thead>
<tr>
<th>Zircon_samples_and_CL_imagery.xls</th>
<th>Spreadsheet file containing zircon sample photos, see Section 3.4. Annotated cathodoluminescence (CL) images for igneous and detrital zircon grains; see Section 3.5.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zircon_U-Pb_Lu-Hf_data_tables.xls</td>
<td>Spreadsheet file containing igneous and detrital zircon U-Pb and Lu-Hf results data tables, Terra-Wasserburg diagrams, see Sections 3.5 and 3.6.</td>
</tr>
<tr>
<td>Figure_3-10_data_table.xls</td>
<td>Contains the compilation of U-Pb, Lu-Hf, and Sm-Nd age and isotopic data from the Kalahari Craton as well as the complete reference list, see Section 3.7.2 and Figure 3.10</td>
</tr>
</tbody>
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