STRATIFORM AND VEIN-TYPE DEPOSITS IN THE PAN-AFRICAN OROGEN IN CENTRAL AND SOUTHERN AFRICA: EVIDENCE FOR MULTIPHASE MINERALISATION

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(7 figures)

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ABSTRACT. Stratiform deposits in the Pan-African Orogen are of the Cu-Co type and are restricted to carbonates and siliciclastic sediments that are stratigraphically close above the basement. The stratiform Cu-Co deposits formed during early diagenesis (possibly around 820 Ma) and during late diagenesis/metamorphism and the Pan-African Orogeny (~580 to ~520 Ma). The early diagenetic Cu-Co sulphides were partly remobilised into the second stratabound Cu-Co mineralisation, with precipitation of Cu-Co sulphides in nodules, veins and as breccia cements.
Vein-type Cu-Pb-Zn mineralisation occurs at two distinct levels, higher in the stratigraphy. The lower level vein-type deposits occur in dolomite and are dominantly of the Zn-Cu type. The higher level vein-type deposits occur at the contact between dolomite and sandstone and are dominated by Cu. The Cu-dominated deposits that have been dated, developed during the waning stage of the Pan-African Orogeny (~530 to ~500 Ma). The Zn-dominated deposits for which a mineralisation age has been established, formed after the Pan-African Orogeny. Some of these vein-type mineralisation have been remobilised after their formation, with the precipitation of massive Cu(-Ag) sulphides.

KEYWORDS: Central and Southern Africa, Neoproterozoic mineralisations, Pan-African Orogen, Stratiform Cu-Co ore deposits, Vein-type Zn-Cu-Pb-Ag deposits,

1. Introduction

The Neoproterozoic rocks of the Pan-African Orogen in Central and Southern Africa (Fig. 1) host numerous large and rich base metal ore deposits of stratiform, stratabound and vein-type. Tenke Fungurume, Luiswishi and Konkola are typical stratiform ore deposits of the Central African Copperbelt in both the Democratic Republic of Congo (here after DRC) and Zambia (Fig. 2A). Kipushi and Dikulushi in the DRC, Kabwe in Zambia and Berg Aukas and Tsumeb in Namibia are examples of stratabound or vein-type deposits (Figs. 2A&B). Several metallogenic models have been proposed for the origin of the stratiform deposits in the Copperbelt (review in Cailteux et al., 2005 and Selley et al., 2005). However, during these last years, results from several studies indicated mineralisation formed from early to late diagenetic processes and from syn-orogenic processes (e.g. Cailteux et al., 2005; Selley et al., 2005; Dewaele et al., 2006c; Muchez et al., 2008). The typical vein-type deposits have a syn- to post-orogenic origin and are also characterised by multiple mineralisation/ remobilisation stages (Heijlen et al., 2008; Haest et al., 2009). In a preliminary study, Muchez et al. (2007) have shown that the different mineralisation and remobilisation stages are related to the geodynamic evolution of the Lufilian Arc.

Recently Kampunzu et al. (2009) discussed in detail the geological setting and mineralogy of the most important sediment-hosted (i.e. vein-type and stratabound) Zn-Pb-Cu deposits in the DRC, Zambia and Namibia. The aim of this paper is to review the most fundamental data of typical deposits in order to present a metallogenic model framed in the geodynamic evolution.

2. Geological setting

The stratiform and vein-type Cu(-Co) ore deposits described above are hosted by Neoproterozoic (meta-) sedimentary rocks of the Pan-African Lufilian and Damara Orogenic Belts (Fig. 1). These sedimentary rocks were deposited in an intracontinental rift followed by a proto-oceanic rift system that finally evolved into a syn- to postorogenic sequence (e.g. Kampunzu et al., 1991; Miller et al., 2009). The stratiform Cu-Co deposits and the vein-type deposits of Kipushi (DRC) and Kabwe (Zambia) are situated in the Lufilian Arc (Fig. 2A). The vein-type deposit of Dikulushi (DRC) is located on the Kundelungu Plateau to the north of this orogenic belt (Fig. 2A). The Berg Aukas-type Zn-Pb stratabound ore deposits and the Cu-dominated Tsumeb-type deposits are situated on the Otavi Mountain Land (here after OML), forming the northern foreland of the Damara Belt in Namibia (Fig. 2B, Pirajno & Joubert, 1993)
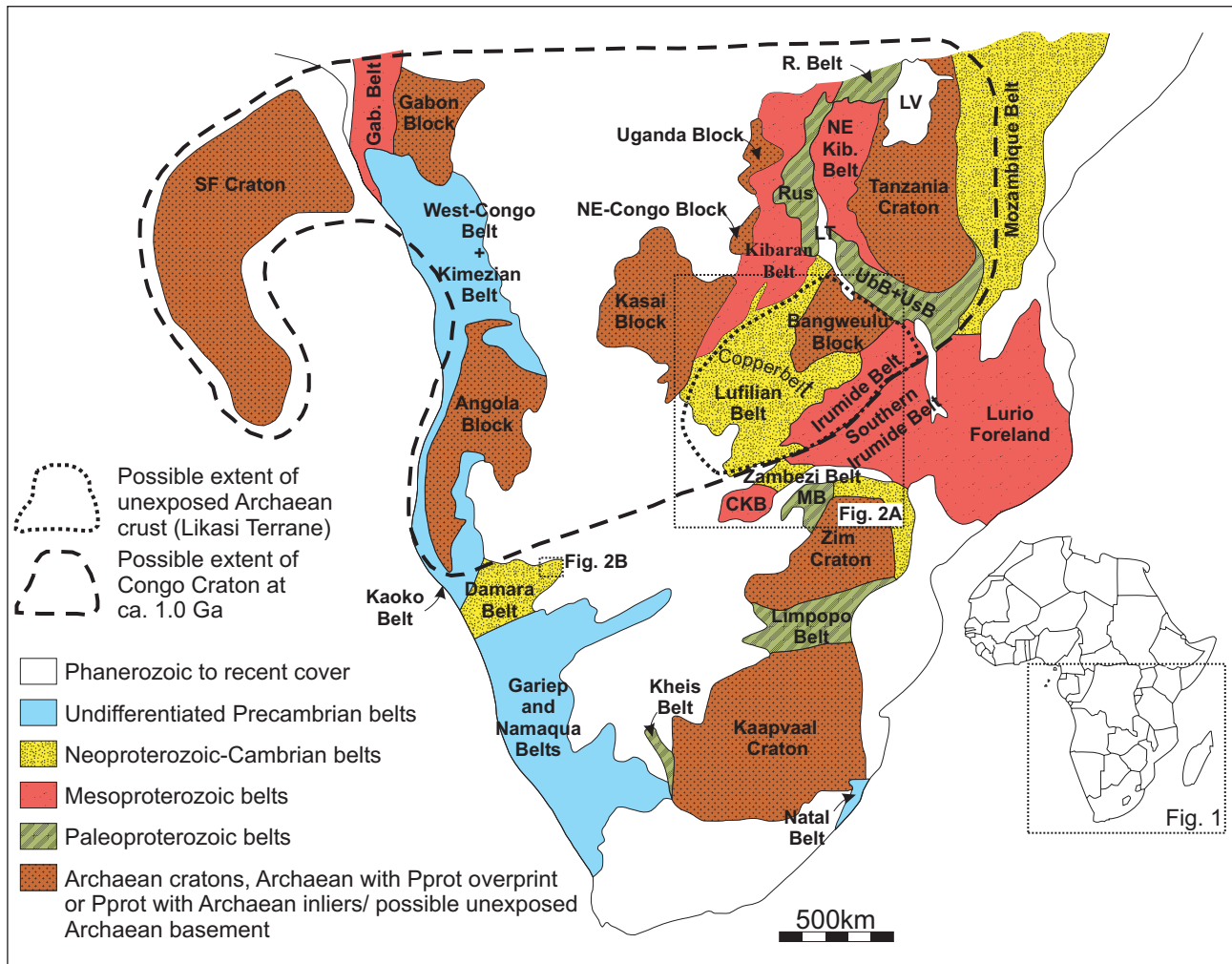
Figure 1. Simplified geological map of Sub-Saharan Africa (map of Africa in the lower right corner outlines the region that is shown. CKB: Choma Kalomo Block; Gab. Belt: Gabon Belt; LT: Lake Tanganyika; LV: Lake Victoria; MB: Magondi belt; NE Kib. Belt: Northeast Kibaran Belt; R. Belt: Ruwenzori Belt; Rus.: Rusizian; SF Craton: Sao Francisco Craton; UbB: Ubendian Belt; UsB: Usagaran Belt; Pprot: Palaeoproterozoic; after De Waele et al., 2008)

The Lufilian Belt is bordered by the Irumide Belt to the east, the Kasai Block to the west, the Kibaran Belt to the north and the Bangwelu block to the northeast (Fig. 2A). The Bangweulu Block includes a basement of granitoids, overlain by a continental clastic succession that contains minor volcanic intercalations near the base (De Waele et al., 2008). The Irumide Belt is characterised by a metasedimentary succession of shallow marine quartzites and pelites that was deposited after 1.8 Ga according to U-Pb dating on detrital zircon (De Waele and Mapani, 2002). Granitoids with a crust-dominated geochemical character intruded the Irumide Belt at 1664 ± 4 Ma and at 1551 ± 33 Ma (De Waele et al., 2006). The southern part of the Kibara Belt consists of at least two amphibolite grade meta-sedimentary successions. The oldest succession was intruded by I-type plutons at 1.38 to 1.37 Ga and deformed (Kokonyangi et al., 2004). A second tectonic event affected both sedimentary successions and is dated at 1089 ± 14 Ma based on a U-Pb zircon SHRIMP age of a syn-kinematic granitoid (Kokonyangi et al., 2004).

The (meta-)sedimentary succession that builds up the Lufilian Belt is called the Katanga Supergroup and is subdivided in the Roan, Nguba and Kundelungu Groups (Fig. 3; Cahen, 1954; Batumike et al., 2007; Cailteux et al., 2007). Deposition of the Katanga Supergroup started after intrusion of the Nchanga granite, which formed at 883 ± 10 Ma (zircon U-Pb SHRIMP: Armstrong et al., 2005). The Roan Group is separated from the Nguba by a conglomerate (Mwale Formation), which is related to the global Sturtian glacial epoch (Batumike et al., 2006). The Nguba is separated from the Kundelungu by another conglomerate (Kyandamu Formation), which is related to the Marinoan ice age (Batumike et al., 2006). The Kundelungu sedimentary rocks represent pre- to post-orogenic sediments (Kampunzu & Cailteux, 1999).

The Damara Belt is located between the Congo (to the north) and Kalahari (to the south) cratons, which both consist of >1100 Ma rocks (Fig. 1, Frimmel et al., 1996). The Angola Block builds up the Congo craton to the north of the Damara Belt (Fig. 1). The southwestern part of the Angola Block consists of granitic to tonalitic-granodioritic
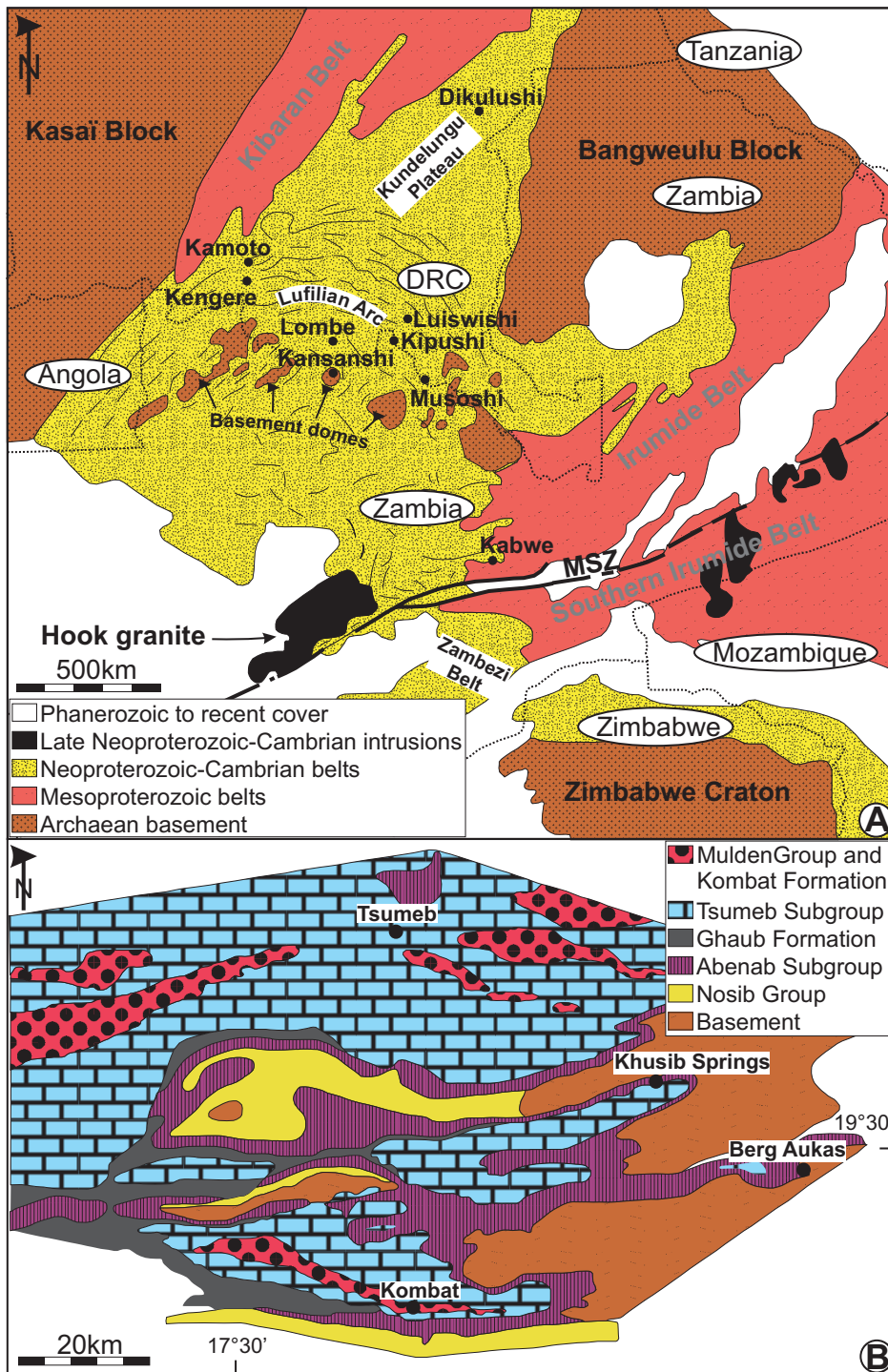
Figure 2. A. Simplified geological map of the Neoproterozoic Lufilian Arc and the Kundelungu Plateau to the north, showing some of the stratiform and vein-type ore deposits in the region (MSZ: Mwembeshi Shear Zone; CKB: Choma-Kalomo Block; modified after Porada and Berhorst, 2000). B. Simplified geological map of the Otavi Mountain Land with location of the Berg Aukas and Tsumeb-type ore deposits (after Söehnge, 1957).

gneisses of 2.5 Ga (Delor et al., 2008). Inherited zircons in adjoining Palaeoproterozoic formations give no ages older than 2.71 Ga. The main part of the Angola Block is covered with siliciclastic formations and interbedded volcanics that have been dated at 2.0 to 1.96 Ga (Delor et al., 2008).

The Damara Supergroup builds up the Damara Belt and is composed of the Nosib, Otavi and Mulden Groups in the OML (Fig. 3; Frimmel et al., 1996). Deposition of the Nosib Group started before ca. 756 Ma (Hoffman et al., 1996). However, no date for the onset of Nosib Group sedimentation is known. By analogy with the similar Stinkfontein Subgroup in the Gariep Belt, Frimmel et al.

(2001, 2004) assumed sedimentation started sometime around 770 Ma, but Frimmel and Miller (2009) argue for the Nosib Group being older and possibly contemporaneous to the Roan Group in the DRC and Zambia. The Nosib Group is separated from the Otavi Group by a conglomerate (Varianto Formation), which is related to the global Sturtian glacial period (Chetty & Frimmel, 2000) and corresponds to the Mwale Formation in the Katanga Supergroup (Fig. 3; Cailteux et al., 2007). The Otavi Group is subdivided in the Abenab and Tsumeb Subgroups that are separated by a conglomerate (Ghaub Formation) that was deposited during the Marinoan ice age (Hoffmann et al., 2004) and can be related to the Kyandamu Formation

	Katanga (DRC)				Namibia			
	Katanga Supergroup				Damara Supergroup			
Age	Group	Subgroup	Formation	Deposit	Deposit	Group	Subgroup	Formation
	Kundelungu	Biano				Mulden		
		Ngule	Sampwe Kiubo Mongwe	Dikulushi	Tsumeb Kombat		Tsumeb	Hüttenberg
								Elandshoek
								Maieberg
		Gombela	Lubudi Kanianga Lusele Kyandamu		Khusib Sp.			Ghaub
635 Ma						Otavi		
	Nguba	Bunkeya	Monwezi Katete	Kipushi Lombe Kengere			Abenab	Auros
								Gauss
		Muombe	Kipushi Kakontwe Kaponda Mwale		Berg Aukas			Berg Aukas
750 Ma								Varianto
	Roan	Mwashia				Nosib		Askevold
		Dipeta			Age?? (>756 Ma)			Nabis
		Mines						
<880 Ma		RAT						

of the Katanga Supergroup (Fig. 3; Cailteux et al., 2007; Kampunzu et al., 2009).

Both the Lufilian and Damara Orogenies occurred between ca 560 and 530 Ma (Hanson et al., 1993; Frimmel & Fölling, 2004; John et al., 2004). Syn-to post-tectonic intrusives and rhyolites in central Zambia were dated by Hanson et al. (1993). Synorogenic granites yielded U-Pb ages of 566 ± 5 Ma and 559 ± 18 Ma. The age of the post-orogenic rhyolite is 538 ± 1.5 Ma. SHRIMP U-Pb data of metamorphic monazite from the Chambishi basin provide ages of 592 ± 22 Ma, 531 ± 12 Ma and 512 ± 17 Ma, which correspond respectively to the ages of eclogite facies metamorphism, high pressure talc-kyanite whiteschist metamorphism, and a regional metamorphic and mineralisation pulse (Rainaud et al., 2005). Post-orogenic cooling possibly started at ca 510 Ma (John et al., 2004; Rainaud et al., 2005). A post-orogenic magmatic pulse in the Gariep Belt (southern Namibia) occurred at 507 Ma (Frimmel, 2000).

Rocks from the Katanga Supergroup close to the core of the Lufilian Belt and from the Damara Supergroup in the OML underwent greenschist facies metamorphism, indicating peak temperatures between 300 and 400 °C were reached in the region (Ramsay and Ridgeway, 1977; Frimmel et al., 1996).

3. Stratiform deposits in the Lufilian Orogen

The aim of this overview is not to provide a detailed review of the metallogenesis of the stratiform Cu-Co ore deposits in the Copperbelt, but to give relevant data and evidence for multiphase mineralisation. Detailed reviews on the Zambian and Katangan Copperbelt are published respectively by Selley et al. (2005) and Cailteux et al. (2005). Recently, a detailed study of the fluid inclusions at Kamoto and Luiswishi (DRC) and a comparison with the microthermometric data from both Zambia and the DRC has been carried out by El Desouky et al. (2009). A study of the C, O, S and Sr isotopic composition of the two main

mineralisation phases in the DRC and a comparison with published data on the Copperbelt has been performed by El Desouky et al. (2010).

The stratiform Cu-Co ore deposits in the Lufilian Orogen occur both in the DRC and Zambia. In the DRC, stratiform Cu-Co mineralisation is hosted in carbonate dominated allochtonous units at two levels of the Mines Subgroup (Cailteux et al., 2005; Fig. 3). Similar stratiform Cu-Co deposits in Zambia are situated in more siliciclastic units that belong to the Mindola and Kitwe Subgroups. These stratiform Cu-Co mineralisation will be discussed together and formed during two main phases, identified in Zambia (Selley et al., 2005) and in the DRC (Cailteux et al., 2005; Muchez et al., 2008; El Desouky et al., 2009). Phase 1 Cu(-Co) mineralisation includes sulphides in bands, in type I nodules and disseminated through the host rock (El Desouky et al., 2010). Phase 2 Cu(-Co) mineralisation includes mineralisation in type II nodules, in veins and as cements in tectonic breccia. Type I and type II nodules are based on the crystal size of the minerals they contain (El Desouky et al., 2010).

Framboïdal and euhedral pyrite are found disseminated in the host-rock around most deposits (Cailteux et al., 2005). Diagenetic pyrite at Nchanga (Zambia) has $\delta^{34}S$ values between -17 and -1 ‰ V-CDT, interpreted to be consistent with low temperature open system bacteriogenic reduction of sea-water sulphate (McGowan et al. 2006). A first generation of dolomite postdates the formation of framboïdal pyrite, but predates the first main phase of Cu(-Co) sulphides (Dewaele et al., 2006c; El Desouky et al., 2009). Phase I Cu(-Co) sulphides include carrolite, chalcopyrite, bornite and chalcocite and occur disseminated, in bands or in nodules type I (El Desouky et al., 2009). The sulphides, dolomite and quartz replace the initial anhydrite in the nodules (Muchez et al., 2008). The nodules are surrounded by organic material. A diagenetic origin for both the stratiform ore deposits in Zambia and the DRC was already proposed in the early 1970's (Annels,
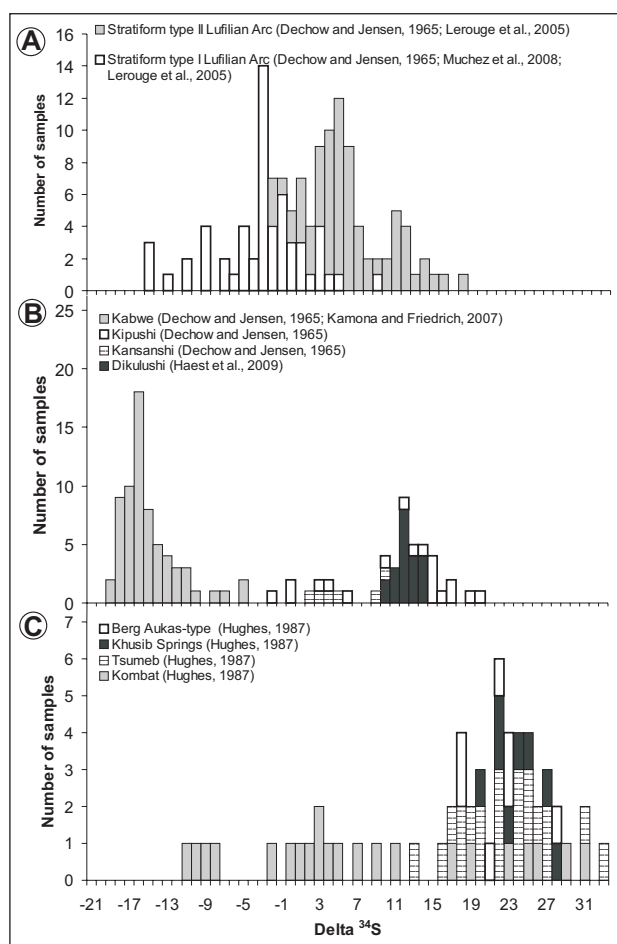
Figure 4. Compilation of the S isotope data of sulphides in stratiform ore deposits in the Lufilian Arc (Stratiform type I: Kamoto (Dechow & Jensen, 1965; Muchez et al., 2008), Luiswishi (Lerouge et al., 2005; Muchez et al., 2008); Stratiform type II: Nkana and Mindola (Dechow & Jensen, 1965), Luiswishi (Lerouge et al., 2005)) (A), in vein-type ore deposits in the Lufilian Arc and on the Kundelungu Plateau (B) and in vein-type ore deposits in the Otavi Mountain Land (C).

1974; Bartholomé et al., 1972; Bartholomé, 1974). An important part formed at the expense of anhydrite, whether this mineral was originally present in nodular form or occurred as an interstitial cement replacing pyrite (Annels, 1974; Bartholomé et al., 1972; Haynes, 1986). Recent Re-Os dating of chalcopyrite at Konkola (Zambia) yielded an age of 816 ± 62 Ma (Selley et al., 2005), which also suggests that Cu(-Co) sulphides precipitated during early diagenesis. At Kamoto (DRC), Luiswishi (DRC), Chambishi (Zambia) and Chambishi Southeast (Zambia), the early diagenetic sulphides precipitated from a moderately saline (11 – 21 eq. wt.% NaCl), low to moderate temperature (Th = 115 – 220 °C) fluid, with a H_2O-NaCl composition (Annels, 1989; El Desouky et al., 2007, 2009). The dolomite in the nodules has $\delta^{18}O$ and $\delta^{13}C$ values of -9.79 to -11.50 ‰ V-PDB and -7.02 to -9.93 ‰ V-PDB respectively, indicating (1) precipitation from a seawater-type fluid at a maximum temperature of 70°C, and (2) incorporation of light ^{12}C, produced during oxidation of organic matter (Muchez et al., 2008). The

source of sulphur would be interstitial dissolved sulphate in the sediments, anhydrite cement or early diagenetic anhydrite nodules (Annels, 1989). Sulphur was reduced through BSR, based on the low $\delta^{34}S$ values (Fig. 4A). However, BSR cannot have occurred during passage of the hydrothermal metal transporting fluid and must have started before the Cu(-Co) mineralisation or could have occurred intermittently between the hydrothermal discharges (Muchez et al., 2008; El Desouky et al., 2010). Oxidation during BSR of the organic material included in the rocks close to the nodules explains the light C isotope composition for the dolomite gangue (Muchez et al., 2008). Furthermore, BSR produced HCO_3^-, which precipitated in dolomite, and H_2S, which precipitated in the associated sulphides (Muchez et al., 2008). A decrease of the pH during sulphide deposition led to the precipitation of quartz, always closely associated to the sulphide mineralisation (Muchez et al., 2008). $^{87}Sr/^{86}Sr$ ratios of dolomite in type I nodules range between 0.71012 and 0.73576 (Muchez et al., 2008), significantly more radiogenic than the strontium isotopic composition of Neoproterozoic marine carbonates ($^{87}Sr/^{86}Sr$ = 0.7056–0.7087, Jacobsen & Kaufman, 1999; Fig. 5A), suggesting intense interaction with siliciclastic sedimentary rocks and/or the granitic basement (Muchez et al., 2008). This
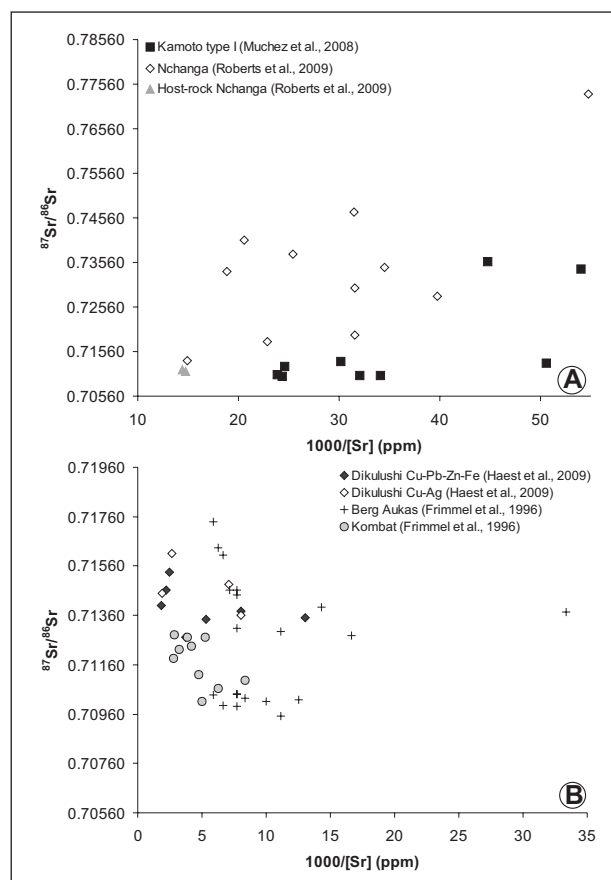

Figure 5. Compilation of Sr isotope data from carbonate gangue minerals associated with early diagenetic nodules in the stratiform deposits of Kamoto and Nchanga in the Lufilian Arc (A) and from carbonate gangue minerals associated with vein-type ore deposits in the Pan African Orogen (B).

supports the hypothesis that gneisses, magmatic rocks and schists from the Palaeoproterozoic basement, observed in the Domes region and in the Zambian Copperbelt, provided Cu to the mineralising fluids (Voet & Freeman, 1972; Ngoyi et al., 1991; Sweeney & Binda, 1994; Cailteux et al., 2005; Dejonghe & Ngoyi, 1995; McGowan et al., 2006; Bernau et al., 2007). The origin of cobalt, the presence of which is not revealed in these basement rocks, was postulated to be sourced from an alkaline magma body at depth (Annels, 1974, 1989; Unrug, 1988) or from the Archean rocks of the Zimbabwe Craton (Cailteux et al., 2005).

A second main phase of Cu(-Co) sulphides is observed in type II nodules, veins and as cements in tectonic breccia (El Desouky et al., 2007, 2009). The main phase II Cu(-Co) sulphides are associated with coarse crystalline quartz and overgrown by dolomite. An identical Cu-Co mineralogy for sulphides from both phases in the DRC indicates that the late stratabound mineralisation probably only remobilised the early stratiform Cu-Co mineralisation in Katanga (Muchez et al., 2008; El Desouky et al., 2010).

This remobilisation has been related to metamorphism and deformation both in Zambia and the DRC (Annels, 1989; Cailteux et al., 2005; El Desouky et al., 2009), as confirmed by radiometric dating at Musoshi (DRC), Nkana, Chibuluma and Nchanga (Zambia), revealing mineralisation ages of 583, 526 and 514 Ma (Richards et al., 1988a, b; Barra et al., 2004). These ages span the entire Lufilian Orogeny and could all indicate individual remobilisation events of early diagenetic Cu(-Co) mineralisation into late diagenetic to syn-orogenic Cu(-Co) mineralisation (El Desouky et al., 2009). The age of 583 ± 24 Ma overlaps with the oldest biotite (~586 Ma) and monazite (~592 Ma) ages (Rainaud et al., 2005) and even with the synorogenic magmatic rocks (560 Ma). At Nkana, several successive mineralised vein generations formed during the Lufilian orogeny, i.e. layer parallel veins, irregular croscutting veins and massive veins (Brems et al., 2009). These vein generations respectively formed during the initial phase of basin inversion, the main phase of deformation and a late phase of orogenesis or extensional tensions related to the collapse of the
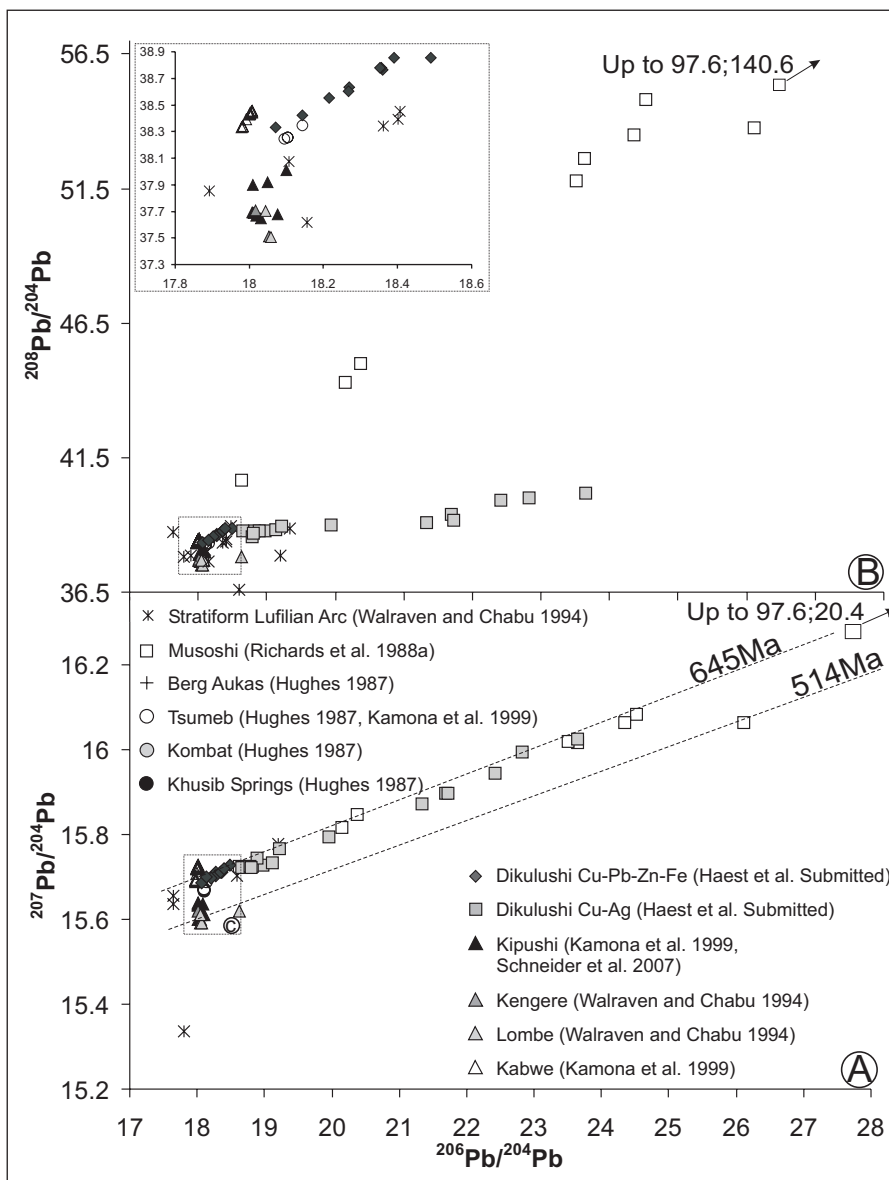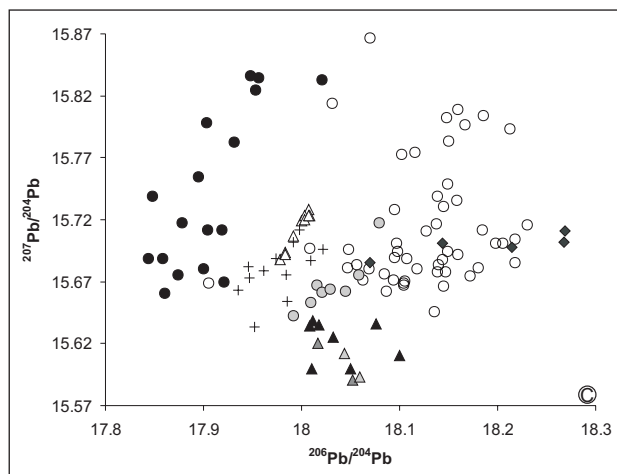

Figure 6. Uranogenic and thorogenic plots for stratiform, stratabound and vein-type deposits in the DRC, Zambia (Kabwe) and Namibia (Berg Aukas, Tsumeb, Kombat and Khusib Springs) (A & B); C Detail of less radiogenic Pb isotope data in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space, indicated with dotted box in (A) (stratiform Lufilian Arc: Kolwezi and Kinsenda; after Kampunzu et al., 2009)

orogen. Late hydrothermal alteration and veining at Musoshi is dated at 514 Ma ± 2 Ma, based on U-Pb rutile dating and at 514 ± 3 Ma, based on U-Pb uraninite dating (Richards et al., 1988a). Richards et al. (1988a, b) linked this hydrothermal phase to compressional deformation and metamorphism during the Lufilian Orogeny or it could be related to extensional tensions as a consequence of collapse of the orogen.

The remobilised Cu(-Co) sulphides in the DRC precipitated from a fluid with a high salinity between 35 and 45.5 eq. wt% NaCl, at a temperature between 270 and 385 °C (El Desouky et al., 2009). In Zambia, fluid inclusions in late stage veins at Chambishi show a broad range in homogenisation temperatures ranging from 200 to 430°C (Annels, 1989). These fluids have an aqueo-carbonic composition with a dominance of CH_4 in certain types (Greyling et al., 2005). Greyling (2009) proposed the early orogenic fluid inclusions were trapped near hydrostatic conditions at a maximum temperature and pressure of 260°C and 800 bar. In an extensive study of the oxygen and carbon isotopic composition of whole rocks, veins and nodules, Selley et al. (2005) found a clear trend in the isotopic signature, which has also been observed in earlier studies (Annels, 1989; Sweeney & Binda, 1989). The isotope data range from values typical of Neoproterozoic marine carbonates ($\delta^{13}C$ = -4 to +4 ‰ V-PDB and $\delta^{18}O$ = -8 to -4 ‰ V-PDB) in the non-mineralised host rock to isotopically light values ($\delta^{13}C$ = -26 to 4- ‰ V-PDB and $\delta^{18}O$ = -25 to -8 ‰ V-PDB) in carbonates associated with the sulphide mineralisation. A first and extensive study on the S isotopic composition of the sulphides in the Copperbelt in Zambia and the DRC was carried out by Dechow and Jensen (1965). About 550 sulphides and 10 sulphates were isotopically analysed. S-isotopic values range between -18 and +23 ‰ V-CDT (data from Nkana and Mindola have been incorporated in Fig. 4A). Based on the quite similar $\delta^{34}S$ values of stratiform sulphides and late stockwork sulphides at Luiswishi (DRC), Lerouge et al. (2005) suggested an in situ recrystallization or a slight remobilisation of stockwork sulphides without attainment of isotopic equilibrium between different sulphide phases. At Nkana (Zambia), an increase of the sulphur isotopic composition

occurs in the successive vein generations that formed during the Lufilian orogeny, going from -29 to +15 ‰ V-CDT, indicating the dominance of remobilised early diagenetic sulphides in the early veins and an increasing contribution of newly formed S by TSR of dissolved Neoproterozoic sulphates in the later vein generations (Muchez et al., 2010).

At Nchanga (Zambia), which is interpreted to be a syn-orogenic Cu-Co deposit (McGowan et al., 2003), dolomites from the hanging wall show $^{87}Sr/^{86}Sr$ values from 0.711 to 0.7116, which are slightly enriched compared with Neoproterozoic seawater water (Fig. 5A; Roberts et al., 2009). Carbonates associated with the Cu-Co sulphides show more radiogenic $^{87}Sr/^{86}Sr$, ranging between 0.7136 and 0.7469 (Fig. 5A; Roberts et al., 2009). Also at Nkana, the strontium isotopic composition of the syn-orogenic veins shows high $^{87}Sr/^{86}Sr$ ratios up to 0.75407 (Muchez et al., 2010). These radiogenic values are interpreted to indicate interaction of the fluid that precipitates the vein cements, with the basement or with sediments derived from it.

The Pb isotopic composition of the stratiform Cu-Co deposits of Kolwezi, Kinsenda and Musoshi from the CAC has been measured (Richards et al., 1988a; Ngoyi et al., 1993; Walraven & Chabu, 1994). Richards et al. (1988a) divided the measured samples in groups based on re-crystallisation textures, grainsize and position (e.g. along veins). The strongly re-crystallised Cu(-Co) sulphides and the Cu(-Co) sulphides that have been sampled from veins that crosscut the stratigraphy might belong to the second generation Cu(-Co) sulphides described above. Walraven & Chabu (1994) did not make any distinction and the significantly larger grain size of the second generation Cu(-Co) sulphides could indicate that they only measured their Pb isotopic composition. The Pb isotopic signature of the stratiform Cu(-Co) sulphides from the CAC varies significantly (Fig. 6). The stratiform sulphides from Musoshi display a linear trend towards highly radiogenic Pb isotopic compositions (Fig. 6A). Richards et al. (1988a) interpreted this linear trend as a Pb-Pb isochron corresponding to an age of 645 ± 15 Ma (Pb-Pb step-leaching of non-recrystallised sulphides; Richards et al., 1988a) that was reworked during uraninite formation around 514 ± 3Ma (U-Pb dating of rutile and uraninite; Richards et al., 1988a, b; Fig. 6A). At Kinsenda, an isochron age of 725 Ma was based on Pb isotopic data of sulphides and interpreted to reflect remobilisation of the primary ore (Ngoyi et al. 1993). Kamona et al. (1999), however, interpreted the linear trend at Musoshi as the result of mixing of Pb from different sources during Lufilian metamorphism. In the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space, the samples from the Musoshi deposit have ^{208}Pb enriched values and are likewise linearly correlated (Fig. 6B). The Musoshi mineralisation is hosted in shales that typically have high Th/Pb ratios, explaining the ^{208}Pb enriched values. The U-Th-Pb system for the Musoshi sulphides appears to be coupled and the 645 Ma age might be considered consistent with respect to the local geology.

Ngoyi et al. (1991) proposed on the basis of a combination of strontium and radiogenic lead isotopes on zircon that the Copperbelt granites at Kinsenda behaved as an open geochemical system during the Pan-African orogeny.

4. Vein-type deposits in the Lufilian Orogen

4.1 Kipushi

The Kipushi deposit is located near the southern border of the Democratic Republic of Congo, with Zambia (Fig. 2A). The mineralisation is situated in the northern limb of a northwest-trending anticline, with an axial plane slightly inclined to the south (see Kampunzu et al., 2009 and references therein). Both limbs of the anticline show concordant successions from the Mwashya Subgroup, over the Muombe Subgroup, to the Bunkeya Subgroup (Fig. 3). The Muombe Subgroup is subdivided in the Mwale, Kaponda, Kakontwe and Kipushi Formations (Batumike et al., 2007). The Kaponda, Kakontwe and Kipushi Formations reach a thickness of 600 m at Kipushi and consist entirely of dolomite, although these formations are normally characterised by an intercalation of thick limestone beds and thin dolomite horizons. The Kipushi Formation consists of gray to black dolomite, intercalated with some black cherts. The rocks are fine-grained and C-rich near the upper contact with the overlying Katete Formation at the base of the Bunkeya Subgroup (Fig. 3). The carbonaceous material is identified as shungite: an amorphous, ungraphitised organic substance. Shungite is present as angular fragments in the sulphide ores and in the barren rocks, but is also frequently observed in globular forms (Francotte & Jedwab, 1963) and infilling fissures and pores (de Magnée & François, 1988). This fact strongly suggests that at least part of the carbonaceous matter has been in liquid form (de Magnée & François, 1988). The sulphides partly replace the shungite (Francotte & Jedwab, 1963). The overlying Katete Formation consists of an alternation of dolomites and dolomitic shales in the Kipushi region (de Magnée & François, 1988). The host-rock of the Kipushi deposit underwent regional, low-grade greenschist facies metamorphism.

The core of the Kipushi anticline consists of a breccia of silicified, talc-rich dolomites in a chloritic silty matrix, mainly derived from older Roan Group rocks, probably belonging to the Dipeta Subgroup (Briart, 1948; Intiomale & Oosterbosch, 1974; Intiomale, 1982; de Magnée & François, 1988; Kampunzu et al., 2009). The axial breccia is bordered by steeply dipping faults and de Magnée & François (1988) propose it is a collapse breccia resulting from dissolution of an evaporitic diapir. A fault with an orientation of 295°/80° to 295°/55° cuts the northern flank of the anticline orthogonally and is called the Kipushi Fault. The Kipushi Fault contains a breccia and separates the concordant, 70° to 80° N dipping regular lithologic succession to the east from the axial breccia to the west. A large slab of stratified shales, silty shales and sandstones with occasional irregular bedding, belonging to the Mongwe and Kiubo Formations (Fig. 3), is located in the hanging wall of the Kipushi Fault and is called the Grand Lambeau (de Magnée & François, 1988). This slab would have sunk in the axial breccia during diapir collapse (de Magnée & François, 1988).

The Kipushi mineralisation is located along the Kipushi Fault, at the contact between the Kipushi and Katete Formations and in pipes that crosscut the Kakontwe and Kipushi Formations (Heijlen et al., 2008; Kampunzu et al., 2009). The Kipushi Fault and the contact between the Kipushi and Katete Formations formed the two major structural features that controlled the position of the mineralisation, with the highest Cu and Zn grades at the intersection of both (Heijlen et al., 2008). The position of the Kipushi deposit along a fault that crosscuts a WNW–ESE-oriented anticline indicates that the mineralisation developed during or after the Lufilian Orogeny. Recently obtained concordant Rb-Sr and Re-Os ages of 451 Ma on sphalerite-bornite and bornite-renierite assemblages from the main sulphide mineralisation indicate it formed ca. 50 Ma after the Lufilian Orogeny (Schneider et al., 2007). See Kampunzu et al. (2009) for a more detailed description and overview of the geology of the Kipushi deposit.

The main sulphide mineralisation at Kipushi overgrew three generations of hydrothermal dolomite and a generation of coarse-crystalline quartz. It is followed by a minor generation of sphalerite. The main sulphide mineralisation replaced the dolomite host-rock and its cements (Intiomale & Oosterbosch, 1974), which commonly resulted in banded ore (Heijlen et al., 2008). Pyrite and arsenopyrite precipitated at the start of the main sulphide mineralising stage and they can be observed as relics in a groundmass of sphalerite and chalcopyrite (Heijlen et al., 2008), together with inclusions of tennantite, germanite, briartite and renierite. Sphalerite shows chalcopyrite disease and smooth interlocking grain boundaries between sphalerite and chalcopyrite indicate that they are largely cogenetic. Bornite and chalcocite precipitated at the end of the main sulphide mineralisation and they selectively replace chalcopyrite (De Vos et al., 1974; Heijlen et al., 2008; Kampunzu et al., 2009). One quartz generation formed contemporaneously with sulphide deposition, as indicated by the common occurrence of euhedral quartz prisms in various sulphides and, conversely, sulphide inclusions in quartz. A dolomite formed after the main sulphide mineralisation and is overgrown by younger, minor pyrite-chalcopyrite-sphalerite mineralisation. The Cu-mineralogy changes from a cementing mass of chalcocite in the upper 100 m to chalcopyrite at depth, following this scheme: chalcocite → chalcocite–bornite → bornite → bornite–chalcopyrite → chalcopyrite (Intiomale & Oosterbosch, 1974). In general, Zn and Pb are preferentially hosted in impure carbonate rocks, while Cu mostly occurs in fractured rocks along the contact with the overlying Katete Formation (Heijlen et al., 2008; Kampunzu et al., 2009).

The main sulphide mineralisation at Kipushi, precipitated from halite-saturated, aqueous NaCl-KCl-$CaCl_2$ brines (ca. 30-43 wt% NaCl equiv), with probably excess carbonic vapour, based on decripitation of fluid

inclusions upon heating before total homogenisation (Heijlen et al., 2008). Molar Cl/Br ratios for the mineralising fluid are close to or higher than the seawater value and therefore the fluid likely acquired its high salinity by dissolution of evaporites (Heijlen et al., 2008). These evaporites could be the diapirs' halite, as proposed by de Magnée & François (1988). Large amounts of fluid are needed to dissolve the diapirs' halite (de Magnée & François, 1988). The origin of the salt-leaching fluids could be formation water or metamorphic fluids that were expelled late in the orogenic evolution (Heijlen et al., 2008), since the Kipushi deposit formed significantly later than the peak of the Lufilian deformation and metamorphism in the region (Schneider et al., 2007). Thoreau (1928) and Intiomale & Oosterbosch (1974) proposed a magmatic hydrothermal origin for the mineralising fluid, related to deep-seated igneous intrusions. Magmatic activity around 450 Ma has been observed in the domes region to the south of Kipushi, where hypabyssal syenite intrusions were emplaced around 458 to 427 Ma (U-Pb zircon ages; Cosi et al., 1992). Also Kampunzu et al. (2009) related the formation of the Kipushi deposit to extension associated with the emplacement of these syenite intrusions. However, except for some gabbro fragments in the axial breccia at Kipushi, no magmatic rocks are associated with the mineralisation.

Whatever the source of the mineralising fluid was, it must have been a high temperature, saline brine, since the main stage mineralising fluid had a temperature of ca. 280-380 °C during mineralisation, based on chlorite geothermometry (287-331°C; Chabu, 1995), sphalerite geothermometry (290-380°C; Ottenburgs, 1964) and based on the molar K/Na content of single inclusions in ore gangue, determined by Heijlen et al. (2008), using LA-ICP-MS (281-353°C). The close agreement of the mineralisation temperature, based on the molar K/Na content of single fluid inclusions, with the temperatures obtained by chlorite and sphalerite geothermometry indicate the major cationic content of the mineralising fluid attained equilibrium with albite and K-feldspar at the temperatures of ore formation (Heijlen et al., 2008).

Sulphides from the discordant orebody along the length of the Kipushi Fault have $\delta^{34}S$ values between +12.8 and +19.2 ‰ V-CDT (Dechow & Jensen, 1965; Fig. 4B). These sulphides are similar to sulphate derived from Neoproterozoic seawater (+10 to + 46 ‰ V-CDT; Claypool et al., 1980; Gorjan et al., 2000; Hurtgen et al., 2002, 2005) and can be explained by TSR (Machel, 1987) or by BSR in a closed system (Ohmoto & Rye, 1979). However, high mineralisation temperatures in excess of 280 °C exclude BSR during precipitation (Machel, 2001) and the narrow isotopic range, close to the isotopic composition of the proposed source sulphate, is much more typical of TSR (e.g. Ohmoto & Rye, 1979; Powell & Macqueen, 1984; Naylor et al., 1989). Sulphides from the concordant orebody along the contact between the Kipushi and Katete Formations have lower $\delta^{34}S$ values between -2.6 and +18 ‰ V-CDT (Fig. 4B). These lower values

indicate S was probably at least partly produced by BSR. The barren dolomitic shales above the mineralised horizon have uniform and lighter $\delta^{34}S$ values between -5.7 and +2.0 ‰ V-CDT (Dechow & Jensen, 1965), also indicating the influence of BSR. Mixing of S from barren dolomitic shales, produced by BSR and likely diagenetic in origin, with S produced by TSR during the main sulphide mineralisation, could explain the lower $\delta^{34}S$ values observed for sulphides from the sub-concordant orebody along the contact between the Kipushi and Katete Formations.

The Pb isotopic composition of the Kipushi deposit plots in a narrow cluster to low radiogenic values (Fig. 6). The Kipushi deposit is located along a fault that crosscuts the Kakontwe and Kipushi dolomites, above a thick sedimentary pile of the Roan Group. Vaasjoki & Gulson (1986) observed that deposits underlain by thick sedimentary piles might be characterised by more homogeneous Pb isotopic signatures due to prolonged mechanical and geochemical reworking and lateral transportation into zones favourable for ore deposition, e.g. carbonate rocks. This could indicate Pb and potentially other metals were probably derived from a vast area of deep-seated rocks, including the Lower Roan Group and the weathered part of the old basement (de Magnée & François, 1988). Interaction of a highly saline fluid with unaltered felsic magmatic rocks at higher temperatures would explain the relatively high Ba, Zn, Fe and Pb contents, compared with metal-bearing brines in similar metamorphic settings (Heijlen et al., 2008). The high Ba content of the mineralising fluid and the absence of baryte in association with the mineralisation at Kipushi indicate no SO_4^{2-} was present in the mineralising fluid or in the host-rock. Also the reduced S content of the mineralising fluid must have been limited by its high metal content. Therefore an ancient reservoir of reduced S is invoked at the site of the current mineralisation (Heijlen et al., 2008).

The main-stage mineralising fluid was sourced from a vast area of deep seated basement rocks and channelised along the Kipushi Fault, with the Grand Lambeau as an impermeable roof. The high barium and ore metal content of the fluid suggests that sulphide precipitation occurred when the metal-bearing fluids encountered a reduced, carbon-rich reservoir that was constantly replenished with H_2S (sour gas) (Heijlen et al., 2008). H_2S was produced by TSR of seawater sulphate, with coeval oxidation of carbonic material (shungite) that is ubiquitous in the sulphides and the host-rock (Heijlen et al., 2008). In the sub-concordant orebody along the contact between the Kipushi and Katete Formations, part of the S was likely derived from mixing of S produced by TSR, with S from the barren dolomitic shales that was produced by BSR and captured in diagenetic pyrites.

The minor pyrite-chalcopyrite-sphalerite mineralisation, post-dating the main sulphide mineralisation, precipitated from a slightly lower saline (ca. 23-31 wt% NaCl equiv) Na-Cl-SO_4 brine at a minimum temperature ranging from ca. 170 to <80°C in

successive fluid inclusion assemblages. These assemblages represent an evolution towards a more oxidised environment with possibly surface-derived meteoric fluids that further dissolved evaporite-bearing strata, based on Cl/Br and Na/Br signatures, higher than seawater, indicating a greater involvement of halite-dissolution brines (Heijlen et al., 2008).

4.2 Kabwe

The Kabwe deposit is located in the central part of Zambia, in the south-eastern extension of the Copperbelt (Fig. 2A), in a region where Neoproterozoic metasedimentary rocks unconformably overlie a Palaeo- to Mesoproterozoic basement. The basement consists of older granite-gneiss and younger Muva schist and quartzite as well as mafic lavas of uncertain age. The metasedimentary cover rocks in the Kabwe region are considered time equivalent to the Katanga Supergroup and are subdivided into the Kangomba and Nyama Formations (Kamona & Friedrich, 2007). The Kangomba Formation consists of a lower member with arkose, quartzite and conglomerate and an upper member with metasiltstone, schist, phyllite and dolomite. The upper member could correspond to the Bancroft Formation in the Zambian Copperbelt (Kampunzu et al., 2009), which is the lateral equivalent of the upper part of the Dipeta and the lower part of the Mwashya Subgroups in the DRC (Cailteux et al., 1994). The Nyama Formation could be equivalent to the Nguba Group or to the upper part of the Roan Group (Kamona & Friedrich, 2007) and consists in the Kabwe area of a matrix supported conglomerate at its base, a finely bedded dolostone alternating with carbonate-rich and carbonaceous shales, and a grey to dark grey bedded dolomite including lenticular carbonaceous shales, and a light grey massive dolomite at the top (see detailed review and discussion in Kampunzu et al., 2009). Rocks from the Kangomba and Nyama Formations underwent greenschist- to amphibolite-facies metamorphism (Ramsay & Ridgeway, 1977).

The stratigraphic position of the massive Pb-Zn mineralisation at Kabwe is unclear (cf. Batumike et al., 2006; Kampunzu et al., 2009). The Kabwe deposit is hosted by massive dolomite, which is correlated by most authors to the upper member of the Kangomba Formation (Brancroft Formation, Upper Roan Group in the Copperbelt), while others correlate these massive dolomites to the massive dolomite at the top of the Nyama Formation (probably Kakontwe Formation, Lower Nguba in the Copperbelt). This uncertainty was not clarified in the most recent papers on the Kabwe deposit by Kamona & Friedrich (2007) and by Kampunzu et al. (2009) who proposed both stratigraphic levels for the Kabwe deposit. Therefore the massive dolomite that hosts the Kabwe deposit is simply called the Kabwe dolomite, without correlation to the stratigraphy of the Katanga Supergroup. The Kabwe dolomite forms part of a large NW–SE-plunging syncline in the Kabwe region. The mineralisation at Kabwe consists of four main, epigenetic, pipe- to vein-like orebodies, controlled by NE–SW oriented faults (Kamona & Friedrich, 2007). All orebodies have associated black organic matter (Kampunzu et al., 2009).

The sulphide mineralisation at Kabwe was preceded by deposition of silica that replaced dolomite (Kamona & Friedrich, 2007). The main sulphide mineralisation started with massive pyrite and was followed by the Ge sulphides, briartite and renierite that developed coeval with chalcopyrite, which contains exsolution intergrowths of both (Kampunzu et al., 2009). The Ge sulphides were replaced by a first generation of sphalerite and galena that can be observed as inclusions in pyrite. Fractures in this pyrite generation were healed by chalcopyrite and the pyrite is overgrown by a second generation of sphalerite (Kamona & Friedrich, 2007). These primary sulphides were deformed, based on cataclastic textures, grain fracturing, lamellar twinning, bent cleavage planes and recrystallisation (Kampunzu et al., 2009). A Pb-Pb model age of ca. 680 ± 13 Ma was provided by Kamona et al. (1999), pre-dating the Lufilian Orogeny, in accordance with the observed deformed sulphides. However, Pb-Pb model ages are not robust and no conclusion should be made about the age of mineralisation at Kabwe (cf. Frimmel et al., 2004).

The second sphalerite generation is commonly overgrown by a second generation of galena that is intergrown with small amounts of covellite and chalcocite. These Cu sulphides are believed to be of secondary origin and they selectively replace chalcopyrite and pyrite from the earlier massive sulphide mineralisation (Kamona & Friedrich, 2007). Most primary sulphide mineralisation is observed as cores of massive sulphide ore, surrounded by oxide zones, with willemite, cerrusite, smithsonite, goethite and hematite that selectively replace sphalerite, galena and pyrite. Oxidation of sulphides persists down to the deepest mining levels at 470 m (Kamona & Friedrich, 2007).

Disseminated pyrite in the barren bedded dolostone below the Kabwe deposit has $\delta^{34}S$ values from -7.5 to -5.1 ‰ V-CDT (Fig. 4B). These pyrites are with minimum 15 ‰ depleted in $\delta^{34}S$ relative to Neoproterozoic seawater sulphate (+10 to + 46 ‰ V-CDT; Claypool et al., 1980; Gorjan et al., 2000; Hurtgen et al., 2002, 2005) indicative of BSR, in agreement with a diagenetic origin for the disseminated pyrite. Galena, sphalerite and pyrite from the main sulphide mineralisation have lower, negative $\delta^{34}S$ values of -17.75 ± 0.28, -16.54 ± 0.27 and -15.82 ± 0.25 ‰ V-CDT respectively (Kamona & Friedrich, 2007; Fig. 4B). These sulphides are with minimum 25 ‰ depleted, compared to Neoproterozoic seawater sulphate, also indicative of BSR (Ohmoto, 1986). Comparison of the S isotopic composition of co-existing sulphide triplets from the main sulphide mineralisation yields concordant isotopic temperatures in the range 227 to 356°C, with a probable temperature of 291+/-35°C (Kamona & Friedrich, 2007). Thus the S from the sulphides must have been reduced prior to the main sulphide mineralisation at Kabwe, since BSR only takes place at temperatures below 80 °C (Machel, 2001). The reduced S was probably stored as disseminated pyrite in the massive dolomite host-rock

and had lower $\delta^{34}S$ values, relative to the disseminated pyrite in the barren bedded dolostone below the Kabwe deposit.

Galena samples from the main sulphide mineralisation at Kabwe have a homogeneous, low radiogenic Pb isotope composition of 17.997 ± 0.007 for $^{206}Pb/^{204}Pb$, 15.713 ± 0.010 for $^{207}Pb/^{204}Pb$ and 38.410 ± 0.033 for $^{208}Pb/^{204}Pb$ (Fig. 6A-C; Kamona et al., 1999). Pb isotope data for the Kabwe and Kipushi deposits have a similarly clustered Pb isotopic composition, but Pb isotope data from the sulphide mineralisation at Kabwe have slightly higher $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (Fig. 6A-C). The Kabwe and Kipushi deposits have similar geological settings, along a fault crosscutting a massive dolomite underlain by a thick sedimentary pile. Thus the Pb isotopic homogeneity of the Kabwe sulphides has probably the same origin as for the sulphides from the Kipushi deposit.

The origin of the ore-forming brine at Kabwe is poorly constrained, although Kampunzu et al. (2009) believe that, based on the speculative 680 Ma Pb-Pb model age for Kabwe, the mineralising fluids were the result of dewatering of Katangan sediments during subduction at the onset of the Lufilian Orogeny. Whatever the age of the mineralisation or the source of the mineralising fluid, the mineralisation formed from a high temperature fluid (~ 300°C) that probably acquired its metal content in much the same way as the mineralising fluid at Kipushi, through leaching of metals from suitable basement rocks. When this metal-bearing fluid migrated upward along NE–SW oriented faults, it precipitated its content as sulphides where it encountered significant quantities of reduced sulphur that were liberated from diagenetic pyrites on the site of the current Kabwe deposit.

4.3 Dikulushi

The Dikulushi deposit is situated in the Kundelungu Plateau region in the southeast of the Democratic Republic of Congo, close to the border with Zambia (Fig. 2A). Mineralisation at Dikulushi is hosted in a dolomite-sandstone sequence from the base of the Kundelungu Group (Fig. 3; Haest et al., 2007). The dolomites belong to the Lubudi Formation at the top of the Gombela Subgroup and the sandstones are intercalated with shales and conglomerates that form part of the Mongwe Formation, which is situated at the base of the Ngule Subgroup (Fig. 3; Haest et al., 2007). A linear NNW-SSE oriented zone with shale that deformed in a ductile manner is located in the Lubudi Formation close to the contact with the Mongwe Formation. This zone of shale is related to a detachment at depth, which caused its emplacement in the Lubudi Formation. It caused brecciation of the surrounding parts of the Lubudi Formation and of the contact zone between the Lubudi and Mongwe Formations (Haest et al., 2007). Similar vertical zones of shale are observed in boreholes below the current excavation level (Haest, 2009). On the surface, mineralisation is hosted in E-W- and NE-SW-oriented faults that crosscut the Mongwe Formation and the brecciated contact between the Mongwe and Lubudi Formations (Haest et al., 2007).

At deeper levels, mineralisation follows the borders of the vertical zones with shale (Haest, 2009).

Two kinds of mineralisation can be distinguished at Dikulushi: a Cu-Pb-Zn-Fe mineralisation that is dominantly observed along E-W-oriented faults in the west of the mine and as remnants along NE-SW-oriented faults in the east, and a Cu-Ag mineralisation that remobilised the earlier Cu-Pb-Zn-Fe mineralisation along NE-SW-oriented faults in the east of the mine (Haest et al., 2007). The Cu-Pb-Zn-Fe mineralisation precipitated during the waning stage of the Lufilian Orogeny at around 520 Ma ago and was remobilised afterwards into the Cu-Ag mineralisation (Haest, 2009). Cu-Pb-Zn-Fe mineralisation at Dikulushi is dominated by galena, sphalerite, chalcopyrite, bornite and chalcocite, with associated quartz, dolomite and calcite gangue that precipitated in a reducing environment from a moderate temperature (Th = 90-140 °C), saline (20-25 eq. wt% $CaCl_2$) H_2O-NaCl-$CaCl_2$ fluid. The Cu-Ag mineralisation consists almost entirely of an orthorhombic chalcocite, with a dense cleavage, post-dating barite, calcite and quartz gangue that precipitated in more oxidising conditions from a low temperature (Th ~ 65 °C), low to moderately saline (<3 to >19 eq. wt% NaCl) H_2O-NaCl-KCl fluid (Dewaele et al., 2006b; Haest et al., 2009, 2010a).

Sulphides from the Cu-Pb-Zn-Fe mineralization and chalcocite and barite from the Cu-Ag mineralisation at Dikulushi have similar $\delta^{34}S$ values between +10.0 and +14.1 ‰ V-CDT (Fig. 4B), in the range of Neoproterozoic seawater sulphate (+10 to +46 ‰ V-CDT; Claypool et al., 1980; Gorjan et al., 2000; Hurtgen et al., 2002, 2005). These sulphides and sulphates have a similar S isotopic composition as the sulphides in the discordant orebody at Kipushi. The sulphur isotopic composition for the sulphides and sulphates at Dikulushi is probably also best explained by TSR, based on the narrow isotopic range, close to the isotopic composition of the proposed source sulphate (e.g. Ohmoto & Rye, 1979; Powell & Macqueen, 1984; Naylor et al., 1989).

Calcite and dolomite associated with the Cu-Pb-Zn-Fe and the Cu-Ag mineralisations at Dikulushi have a similar Sr-isotopic composition between 0.71270 and 0.71610 $^{87}Sr/^{86}Sr$ (Fig. 5B; Haest et al., 2009). The $^{87}Sr/^{86}Sr$ ratios of the carbonate gangue minerals are significantly enriched in ^{87}Sr, compared to the strontium isotopic composition of Neoproterozoic marine carbonates ($^{87}Sr/^{86}Sr$=0.7056–0.7087, Jacobsen & Kaufman, 1999), similar to the carbonates in the type I nodules at Kamoto (compare Figs. 5A&B).

Sulphides from the Cu-Pb-Zn-Fe and the Cu-Ag mineralisation at Dikulushi have a clearly distinct Pb isotopic composition. Cu–Pb–Zn–Fe polysulphides have a relatively unradiogenic and internally inhomogeneous lead isotopic composition ($^{206}Pb/^{204}Pb$=18.07–18.49; Figs. 6A-C), most likely generated by mixing of Pb from isotopically heterogeneous clastic sources (Haest et al., 2010b). Chalcocite samples from the Cu-Ag mineralisation contain only trace amounts of lead and show mostly

radiogenic Pb isotope signatures that fall along a linear trend in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram ($^{206}Pb/^{204}Pb$=18.66–23.65; $^{207}Pb/^{204}Pb$ = 15.72–16.02; Fig. 6A; Haest et al., 2010b). These anomalous characteristics reflect a two-stage evolution involving admixture of both radiogenic lead and uranium during a young fluid event possibly c. 100 Ma ago (Haest et al., 2010b). The Pb isotope systematics of local host rocks to mineralisation also indicate some comparable young disturbance of their U–Th–Pb systems, related to the same event. They could have provided Pb with sufficiently radiogenic compositions that was added to less radiogenic Pb remobilised from precursor Cu–Pb–Zn–Fe polysulphides, whereas the U most likely originated from external sources (Haest et al., 2010b).

Pb isotopic data for chalcocite from the Cu-Ag mineralisation are less radiogenic than for sulphides from the stratiform Musoshi deposit, but on an uranogenic plot these chalcocites lie along the 645 Ma isochron of Richards et al. (1988a; Fig. 6A) and give an identical apparent age. In the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space, the samples from the Musoshi deposit deviate towards ^{208}Pb enriched values, but are likewise linearly correlated (Fig. 6B). This deviation can probably be attributed to the host rocks since the Musoshi mineralisation is hosted in shales that typically have high Th/Pb ratios whereas the Dikulushi Cu-Ag mineralisation is hosted by sandstones that have lower Th/Pb. We suggest that the identical ^{207}Pb-^{206}Pb apparent ages for Musoshi and Dikulishi are a coincidence because geological criteria clearly exclude a mineralisation age as old as ca. 645 Ma for the Dikulushi Cu-Ag mineralisation and Haest et al. (2010b) showed that the linear trend for the Cu-Ag mineralisation at Dikulushi can be attributed to an admixture of radiogenic Pb and possibly U at ca. 100 Ma.

4.4 Smaller vein-type deposits

A large number of smaller vein-type deposits occur in the Central African Copperbelt. Most of them are of the Pb-Zn type, including the deposits of Kengere and Lombe in the DRC. The vein-type deposit of Kansanshi in Zambia forms an exception, being of the Cu-Au type. A recent, detailed review of these deposits is given in Kampunzu et al. (2009). The Kengere deposit is located ca. 50 km to the south of Kolwezi and is the westernmost Pb-Zn deposit of the Copperbelt in the DRC (Fig. 2A). The mineralisation at Kengere is hosted in breccias and fractured rocks, associated with a N–S trending fault, at the contact between the Kaponda and Kakontwe Formations (Kampunzu et al., 2009). Both formations are characterised by the occurrence of black amorphous organic matter. The barren Kipushi Formation overlying the Kakontwe Formation is locally affected by fracturing and minor quartz-carbonate precipitation. A major NE–SW trending fault zone occurs to the north of the deposit and contains a large breccia, which is sub-parallel to the axial fold plane (Kampunzu et al., 2009). The primary sulphide mineralisation at Kengere shows a paragenetic sequence from pyrite over dark brown, Fe-Cd-bearing sphalerite to

Ag-rich galena (Intiomale, 1982). Supergene alteration of the sulphide mineralisation produced secondary Zn-Pb-(Cu) oxide minerals, including cerussite, smithsonite, calamine and traces of chrysocolla.

The Lombe deposit is situated in the central-southern part of the Copperbelt in the DRC (Fig. 2A). It is hosted in carbonate units along a tectonic contact between the Kipushi and Katete Formations (Batumike et al., 2006). The Kipushi Formation contains impregnations of black amorphous organic matter. A large NE–SW trending fault is located south of the deposit and contains metabasic rocks from the Dipeta Subgroup (Kampunzu et al., 2009). Mineralisation at Lombe consists of a main orebody with massive Pb-Zn sulphides, concordant to the dip of the host-rock and a lenticular sub-economic zone of disseminated chalcopyrite and bornite mineralisation in the upper beds of the Katete Formation. Precipitation of the main Pb-Zn sulphide orebody started with pyrite, followed by sphalerite, tennantite, an unidentified Pb-As sulphide, galena, chalcopyrite and arsenopyrite (Intiomale, 1982). The main sulphide mineralisation was oxidised with the formation of covellite, calamine, smithsonite, cerussite, malachite and azurite. The Pb isotopic composition of sulphides from Lombe and Kengere plot in a narrow cluster together with sulphides from the Kipushi deposit (Fig. 6), except one sulphide from Lombe that shows a ^{206}Pb enriched value. The deposits of Kengere and Lombe are also located along faults that crosscut the Kakontwe Formation, above a thick sedimentary pile of the Roan Group. As described above for the Kipushi and Kabwe deposits, their homogeneous Pb isotopic signature might be due to prolonged mechanical and geochemical reworking and lateral transportation into zones favourable for ore deposition, e.g. carbonate rocks.

The Kansanshi deposit is located in northern Zambia, 150 km west of the Zambian Copperbelt (Fig. 2A). The Kansanshi deposit is hosted by metasedimentary rocks that belong to the Katanga Supergroup and include quartz-biotite schist, biotite-garnet schist, carbonaceous phyllites, impure marbles, and quartz-muscovite phyllites (Torrealday et al., 2000). The exact stratigraphic position of these rock units is unknown, because they form part of thick-skinned thrust sheets that contain slices of pre-Katangan basement (Coward & Daly, 1984). The metamorphic grade is upper greenschist - lower amphibolite facies. A large sill of metamorphosed gabbro that developed prior to tectonic activity and mineralisation at Kansanshi, is located to the north of the deposit. The Kansanshi deposit consists of three primary high-angle, sheeted, coarse-grained vein sets, with crosscutting relationships that could be used to constrain their relative age. The first set of veins consists of quartz-dolomite-chalcopyrite-pyrite-(molybdenite) minerals, with moderate to intense albitic and ferroan dolomite alteration of schists and phyllites that resulted in the destruction of metamorphic biotite (Torrealday et al., 2000). The second set of veins contains ferroan dolomite-calcite-pyrite-pyrrhotite-(chalcopyrite)-(molybdenite), with ferroan dolomite alteration. Both vein sets are surrounded by

envelopes of disseminated chalcopyrite. The third vein set contains coarse-crystalline calcite-ferroan dolomite–monazite–molybdenite-(brannerite)-(chalcopyrite), without an alteration envelope.

Re-Os dating on molybdenite from the first two vein sets yielded mineralisation ages that overlap within their 2σ uncertainties, indicating that there is no discernable difference in the ages of these veins at the 2σ level. Combination of Re-Os data for both vein sets yielded an age of 512.4 ± 1.2 Ma (Torrealday et al., 2000). The third molybdenite-rich vein set yielded a Re-Os age of 502.4 ± 1.2 Ma, which is consistent with the observed crosscutting relationship observed in the field. These results indicate that the Cu mineralisation at Kansanshi developed in the Cambrian, after the Lufilian Orogeny during two discrete mineralisation pulses at ca. 512 Ma and ca. 502 Ma (Torrealday et al., 2000).

The sulphides at Kansanshi have a S isotopic composition between +2 and +10 ‰ V-CDT, which is lower than the $\delta^{34}S$ values for sulphides from Dikulushi and falls in the lower part of the range for the sulphides from Kipushi (Dechow and Jensen, 1965). The sulphides at Kansanshi have S isotope ratios that are lower than that of the proposed sulphate source, i.e. Neoproterozoic seawater sulphate (+10 to + 46 ‰ V-CDT; Claypool et al., 1980; Gorjan et al., 2000; Hurtgen et al., 2002, 2005). The presence of organic matter in the host-rocks produced ^{13}C-depleted pyrobitumen and a large amount of CO_2 gas (Kribek et al., 2005). The presence of CO_2 gas during mineralisation is confirmed by the observation of two-phase (L+V) CO_2-rich inclusions in quartz gangue, associated with the mineralisation (Kampunzu et al., 2009). Interaction between graphitised organic matter and the hydrothermal fluid likely controlled the oxidation state of the ore fluid (Kribek et al., 2005) and could have been the controlling factor leading to ore precipitation at the current site.

5. Sediment-hosted ore deposits in the Otavi Mountain Land

More than 600 Cu-Pb-Zn-V deposits and occurrences have been recognised in the OML and these are subdivided in two types, based on geological, mineralogical and geochemical data (Frimmel et al., 1996): (1) stratabound Pb-Zn dominated deposits of the Berg Aukas-type are hosted in platform carbonates of the lower Abenab Subgroup, which is interpreted to be time-equivalent to the Bunkeya and Muombe Subgroups in the DRC and (2) Cu-Pb-Zn deposits of the Tsumeb-type are hosted in platform carbonates of the Tsumeb Subgroup, which is time-equivalent to the Gombela and Ngule Subgroups in the DRC (Fig. 3).

5.1 Berg Aukas-type deposits

The Berg Aukas deposit is situated close to the eastern border of the Otavi Mountain Land (Fig. 2B). Berg Aukas-type Zn-Pb-V deposits are stratabound and, in places, stratiform and are situated in the Berg Aukas Formation.

Discordant, brecciated, pipe-like ore bodies extend away from the stratiform ore into the hanging wall. These ore pipes are usually related to thrusts and faults along which karst chimneys were formed by solution collapse (Frimmel et al., 1996). Mineralisation in the stratiform ore bodies and the karstic pipes is dominated by sphalerite and galena, with only very little to no Cu minerals present. Dolomite is the most important gangue mineral. The ores in Berg Aukas-type deposits have been often oxidised, with the formation of willemite, smithsonite, cerrusite and other secondary minerals (Frimmel et al., 1996).

The main sulphide mineralisation precipitated from a saline fluid (23 eq. wt.% NaCl), at a minimum temperature between 100 and 200 °C (Th_{av} = 180 °C; Frimmel et al., 1996). Frimmel et al. (1996) estimate the formation temperature of the main sulphide mineralisation at 240°C, taking the overburden into account. Crush leach analysis of fluid inclusions in dolomite gangue indicate that the main sulphide mineralisation at Berg Aukas precipitated from a fluid, with low Cl/Br and Na/Br values that point to seawater evaporation as a process to acquire a moderate salinity, rather than halite dissolution (Chetty & Frimmel, 2000). The oxygen isotope composition of dolomite associated with the main sulphide mineralisation varies between -9.1 and -5.6 ‰ V-PDB. Frimmel et al. (1996) calculated that this dolomite generation ($\delta^{18}O_{av}$ = -7.1 ‰ V-PDB) must have precipitated from a fluid, with a $\delta^{18}O$-value of 13 ‰ V-SMOW at a temperature of 240°C. This mineralising fluid was strongly enriched in ^{18}O, relative to seawater or meteoric water, which supports an evolved diagenetic/metamorphic origin of the fluid (Frimmel et al., 1996). The $\delta^{13}C$-value for all dolomite gangue that precipitated before, during and after the main-sulphide mineralisation, is similar and varies between 1.3 and 4.5 ‰ V-PDB, in the range of Neoproterozoic seawater dolomite ($\delta^{13}C$ = -4.0 to +4.0 ‰ V-PDB; Veizer & Hoefs, 1976).

Sulphides from Berg Aukas have $\delta^{34}S$ values between 20 and 30 ‰ V-CDT (Fig. 4, Melcher et al., 2006). These $\delta^{34}S$ values fall in the range of $\delta^{34}S$ values for Neoproterozoic evaporites (+10 to + 46 ‰ V-CDT; Claypool et al., 1980; Gorjan et al., 2000; Hurtgen et al., 2002, 2005), which is confirmed by Hughes (1987) who measured $\delta^{34}S$ values of 31 and 33 ‰ V-CDT for gypsum and anhydrite in an evaporite horizon in the Upper Tsumeb Subgroup. The correspondence in S isotopic composition between the sulphides at Berg Aukas and the S isotopic composition of Neoproterozoic evaporites might indicate the sulphides at Berg Aukas derived their S through TSR of sulphate released during dissolution of evaporites in the sedimentary column.

Host-rock dolomites to the Berg Aukas deposit have $^{87}Sr/^{86}Sr$ ratios that vary between 0.7078 and 0.7108 (Frimmel et al., 1996), similar or slightly enriched relative to Neoproterozoic marine carbonates ($^{87}Sr/^{86}Sr$ = 0.7056 - 0.7087; Jacobsen & Kaufman, 1999). The dolomite gangue to the sulphide mineralisation has $^{87}Sr/^{86}Sr$ ratios that vary between 0.7039 and 0.7174 (Fig. 5B), similar or enriched relative to Neoproterozoic marine carbonates,

indicating contamination of the mineralising fluid with [87]Sr by interaction with [87]Sr-rich host-rocks (Frimmel et al., 1996). Pb-isotope data for sulphides from the Berg Aukas-type deposits are slightly heterogeneous and plot at the less radiogenic side of the Pb isotopic cluster defined by the Cu-Pb-Zn-Fe mineralisation at Dikulushi (Fig. 6C), close to the homogeneous Pb isotopic field of the sulphides from Kabwe (Fig. 6C). Melcher et al. (2003) deduced a linear trend from these data, which they attribute to mixing of Pb from a hydrothermal end-member, with Pb from a different source and/or from the host-rocks along which the mineralising fluids were channelised. Frimmel et al. (1996) consider the underlying volcanic rocks of the Askevold Formation (Nosib Group) as the geologically most sensible source of Pb. Their main centre of volcanism is in the south of the OML. They have been metamorphosed to greenschist facies and in places strongly altered as evidenced by intense epidotization and calcitization, indicating flushing of the Nosib Group by basinal brines (Frimmel et al., 1996). The Pb isotope composition of these sources and the host-rocks through which the mineralising fluids migrated have not been determined.

5.2 Tsumeb-type deposits

The deposits of Tsumeb, Kombat and Khusib Springs are the most important Tsumeb-type deposits in the Otavi Mountain Land (Fig. 2B). They are not strictly stratabound, but are often hosted in brecciated zones in massive dolomite from the Maieberg, Elandshoek and Hüttenberg Formations (Fig. 3), with associated carbonic material (Chetty & Frimmel, 2000). These brecciated zones have been identified as karst structures in the deposits of Tsumeb and Kombat, filled with feldspathic sandstone from the overlying Mulden Group (Lombaard et al., 1986). The deposit of Tsumeb formed at ca. 530 Ma, based on the close relation between mineralisation and syn-orogenic structures (Frimmel et al., 1996) and based on a preliminary Re-Os model age of 527 Ma (Melcher et al., 2003). These ages indicate the Tsumeb deposit formed at the end of the Damaran Orogeny (Frimmel et al., 2004).

Mineralisation in the Tsumeb-type deposits is dominated by Cu, with lesser amounts of Pb and Zn and an important enrichment in Ag, As, Ge, Ga and Cd. The hypogene mineralisation assemblage at Tsumeb contains pyrite, followed by chalcopyrite, bornite, germanite, sphalerite and gallite, enargite, tennantite and briartite, renierite and galena. Ore emplacement at Tsumeb post-dated much of the regional folding, but the ores bear the imprint of the waning stages of deformation (Lombaard et al., 1986). At Kombat, bornite precipitated first, followed by chalcopyrite and galena, with subordinate sphalerite, tennantite and pyrite (Frimmel et al., 1996). Footwall mineralisation at Khusib Springs contains sphalerite, chalcopyrite, pyrite and minor tennantite. Massive Cu ore crosscuts the footwall mineralisation at Khusib Springs and consists of early chalcopyrite and bornite, overgrown by pyrite and sphalerite, which were replaced by enargite and tennantite, with Ge-bearing colusite, pearceite-polybasite, galena and some pyrite (Melcher et al., 2006).

Calcite and quartz are the dominant gangue minerals in all Tsumeb-type deposits, with minor dolomite and hardly any fluorite or baryte (Frimmel et al., 1996). The ores were enriched by supergene processes after their formation, with the precipitation of massive ore assemblages that comprise chalcocite, digenite, covellite, cuprite, native copper and silver.

The main phase Cu mineralisation at Kombat, Tsumeb and Khusib Springs precipitated from a fluid, with a salinity between 17 and 23 eq. wt% NaCl at a minimum temperature between 190 and 280 °C (Frimmel et al., 1996; Chetty & Frimmel, 2000). Ga/Ge ratios of sphalerites at Khusib Springs and at Tsumeb indicate mineralisation temperatures of $>260 \pm 10$ °C and close to 300 °C respectively (Melcher et al., 2006). Frimmel et al. (1996) estimated the effective formation temperature of the main sulphide mineralisation in the Kombat and Tsumeb deposits, based on fluid inclusion data and taking the overburden in account, at 450 °C. The formation temperature of the main sulphide mineralisation at Khusib Springs was estimated at 370 °C, also using fluid inclusion data (Chetty & Frimmel, 2000). The effective mineralisation temperature for the Tsumeb-type deposits is taken between 300 and 450 °C. Crush leach analysis of fluid inclusions in quartz, calcite and dolomite gangue indicate that these deposits precipitated from a fluid, with low Cl/Br and Na/Br values, indicative of seawater evaporation as a process to acquire a moderate salinity, rather than halite dissolution (Chetty & Frimmel, 2000).

The calcite gangue in the Tsumeb deposit has δ^{18}O-values between -12.1 and -6.4 ‰ V-PDB, with an average value of -10.0 ‰ V-PDB. If calcite precipitated at a temperature of 450 °C, the mineralising fluid had a δ^{18}O-value of ca. 18 ‰ V-SMOW, which is significantly enriched relative to seawater (Frimmel et al., 1996). This fluid probably represents an evolved diagenetic or metamorphic fluid. The same calcite gangue has δ^{13}C-values between -4.4 and 4.0 ‰ V-PDB, which is in the range of Neoproterozoic seawater dolomite in Northern Namibia (δ^{13}C = -5.0 to +5.0 ‰ V-PDB; Halverson et al., 2005). A calcite generation, pre-dating sulphide deposition at Kombat and a calcite generation post-dating sulphide mineralisation at Tsumeb have δ^{13}C-values down to -7.2 and -9.5 ‰ V-PDB respectively (Chetty & Frimmel, 2000), which could indicate interaction with organic material

δ^{34}S-values for sulphides from Tsumeb and Khusib Springs vary between +17 and +27 ‰ V-CDT and between +21 and +28 ‰ V-CDT respectively (Fig. 4C), similar to the sulphides from Berg Aukas. Sulphides from the Kombat mine cross this range with δ^{34}S-values between -11 to +26 ‰ V-CDT (Fig. 4C; Hughes, 1987). Melcher et al. (2003) relate the high δ^{34}S-values to derivation of S from evaporite beds in the Damara Supergroup. As discussed above for the Berg Aukas deposit, evaporite beds in the Tsumeb Group are indeed characterised by δ^{34}S-values between 31 and 33 ‰ V-CDT. Mobilisation of their SO_4^{2-} and reduction by TSR could explain the observed δ^{34}S-values. The low δ^{34}S-values for sulphides at Kombat, down to -11 ‰ V-CDT, correspond to a S

isotopic fractionation of at least 21 ‰ between S from Neoproterozoic seawater sulphates and sulphides, indicating BSR (Ohmoto, 1986). The BSR reduced S was possibly stored in the Kombat deposit as disseminated, early diagenetic pyrite and dissolved during formation of the main sulphide mineralisation at Kombat to supply the S.

$^{87}Sr/^{86}Sr$ isotope ratio data for the carbonate host-rock of the Kombat deposit range between 0.7076 and 0.7086 (Frimmel et al., 1996), in the range of Neoproterozoic seawater carbonates ($^{87}Sr/^{86}Sr = 0.7056 - 0.7087$; Jacobsen & Kaufman, 1999). Calcite gangue from the main sulphide mineralisation at Kombat is enriched in ^{87}Sr, with $^{87}Sr/^{86}Sr$ ratios of 0.7101 to 0.7141 (Fig. 5B), indicating interaction of the mineralising fluid with ^{87}Sr enriched parts of the basement or with other radiogenic rocks along the flow path of the mineralising fluid. Pb isotope data for sulphides from Tsumeb, Kombat and Khusib Springs are low radiogenic and plot around the Pb isotopic cluster defined by the sulphides from the Dikulushi Cu-Pb-Zn-Fe mineralisation and to lower radiogenic values (Fig. 6C). Melcher et al. (2003) suggest Pb isotopic data for the different Tsumeb-type deposits show mixing trends just like for the Berg Aukas-type deposits that could be related to mixing of Pb from a common hydrothermal end member with Pb from different sources and/or aquifers (Hughes, 1987). A comparison with Pb isotope data from the Rosh Pinah ore province made Frimmel et al. (2004) conclude that the common hydrothermal endmember for the Tsumeb-type deposits likely represents Pb derived from Palaeo- to Mesoproterozoic basement that was mixed with Pb mobilised from the local host-rock by syn-orogenic fluids.

6. Multiphase sediment-hosted mineralisation model for the Pan-African Orogen

Sediment-hosted mineralisation in the Central and Southern part of Africa developed during multiple phases that likely span a period of over 400 Ma. Dolomitisation occurred during early diagenesis in the Roan rift basin and is followed by a first main phase of stratiform Cu-Co mineralisation, dated at ca. 816 Ma (Selley et al., 2005; ca. 820 Ma in Fig. 7). The first main phase of stratiform Cu-Co mineralisation occurs in carbonates and siliciclastics from the Lower Roan (Fig. 3). At ca. 816 Ma, the Lower Roan was buried to not more than 1 km depth (Cailteux et al., 1994), which is still within limits of an early diagenetic environment (600-1000 m in Machel, 1999). As Annels (1974) suggested, early diagenetic Cu-Co mineralisation precipitated from a moderate temperature, moderate saline fluid that interacted with sedimentary rocks, the granitic basement and/or an alkaline magma body at depth (Voet & Freeman, 1972; Annels, 1974, 1989; Unrug, 1988; Cailteux et al., 2005; McGowan et al., 2006; Bernau et al., 2007; Muchez et al., 2008; Fig. 7). Interaction with these different rock units and absence of complete mixing would explain the variable Pb isotopic values of these deposits (Fig. 6). S was partly derived from gypsum or anhydrite, previously present in the evaporite nodules and reduced by BSR (Muchez et al., 2008).

The early main phase of stratiform Cu-Co mineralisation was remobilised by a high temperature fluid in a second Cu-Co mineralisation during late diagenesis/metamorphism and during the Lufilian Orogeny (580 Ma to 520 Ma in Fig. 7). In the DRC, the 2nd main Cu-Co mineralisation occurs in nodules, veins or as breccia cements, within the Mines Subgroup (El Desouky et al., 2007, 2009). In the Zambian Copperbelt, precipitation occurred in multiple, successive vein generations during and possibly after the Lufilian orogeny (Brems et al., 2009).

The vein-type Cu-Pb-Zn deposits of Dikulushi and Tsumeb formed in the front zone of the Lufilian and Damaran Belts respectively, near the end of the Pan African Orogeny (530 Ma to 500 Ma in Fig. 7). The
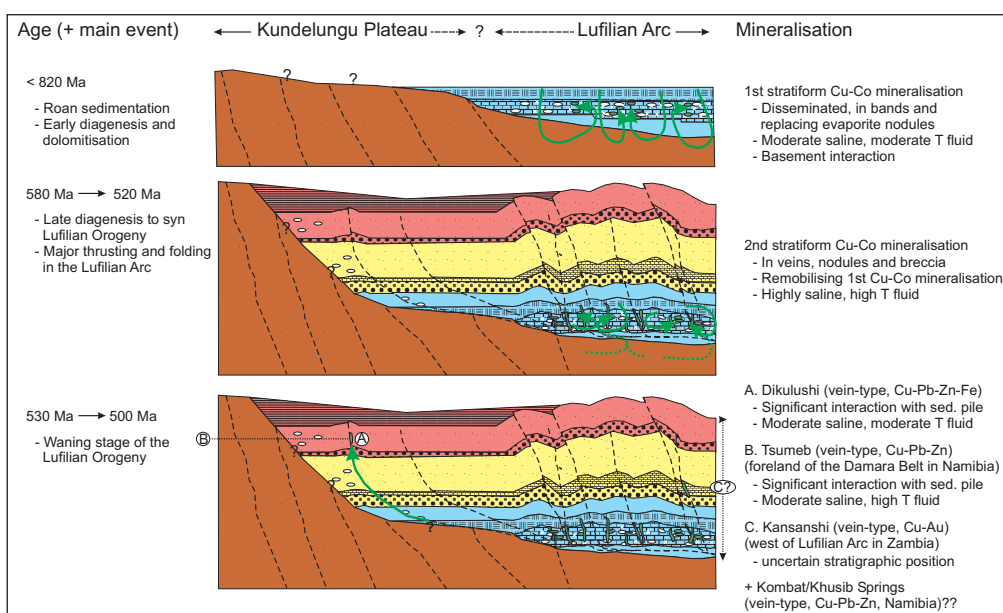

Figure 7. Cu mineralisation in the Lufilian Arc and on the Kundelungu Plateau and its relation to the geodynamic evolution. Reference is made to Cu-Pb-Zn mineralisation in the Otavi Mountain Land (geological cross sections after Dewaele et al., 2006a). 1st and 2nd stratiform Cu-Co mineralisation stages, respectively at <820 Ma and 580 - 520 Ma, modified after El Desouky et al., 2010.
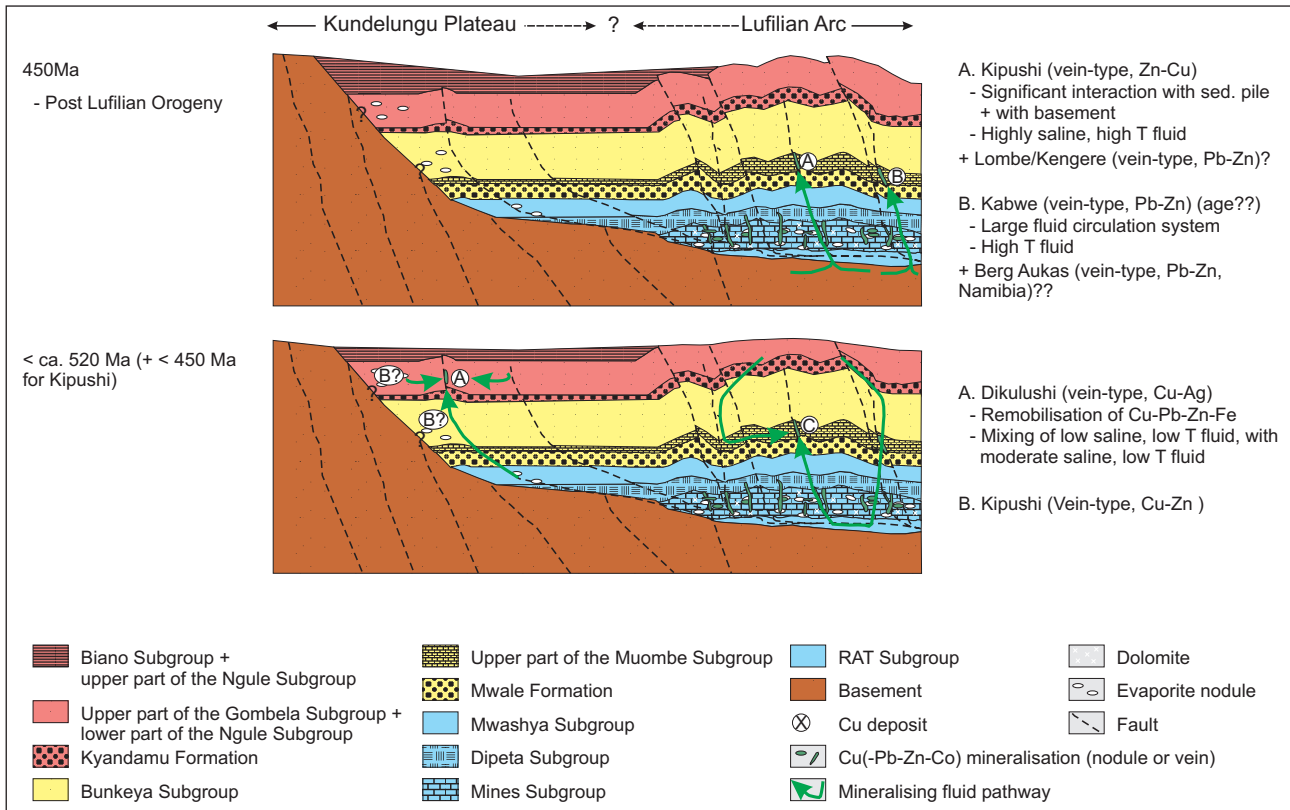
Dikulushi deposit is situated at the level of the Gombela and Ngule Subgroups in Katanga and the Tsumeb deposit is situated at a stratigraphically equivalent position in Namibia. Mineralisation at both localities precipitated from moderate saline, moderate to high temperature fluids that interacted with the basement and the sedimentary pile below. Limited hydrothermal mixing produced internally heterogeneous low radiogenic clusters in the ^{207}Pb/^{204}Pb vs. ^{206}Pb/^{204}Pb space. The sulphides from Dikulushi and Tsumeb might have acquired their S from late Neoproterozoic evaporite layers at depth or from infiltrated seawater. The presence of Neoproterozoic evaporite layers below the Kundelungu Plateau is however uncertain. S in both deposits was reduced by TSR. The Kombat and Khusib Springs deposits in the OML show many similarities with the above deposits, which could indicate a comparable origin. They are all situated at more or less the same stratigraphic level and their mineralogy is dominated by Cu, with minor amounts of Pb and Zn in dolomite-rich lithologies. Their Pb isotopic composition is internally heterogeneous and forms a ^{207}Pb-enriched cluster. However, S in the Kombat deposit was reduced by BSR and likely stored on site in the form of early diagenetic disseminated pyrite that was dissolved during the main sulphide mineralisation. The vein-type Cu-Au deposit of Kansanshi developed in the same period, but its stratigraphic position is uncertain and it contains Au in addition to Cu, rather than Pb and Zn.

These vein-type deposits might have formed together or after the second main phase of Cu-Co mineralisation in the Lufilian Arc. Except for their partly overlapping age and the fact that S in most of the sulphides from these deposits was the result of TSR, they have little in common. The vein-type deposits all formed in the front zone of the Pan-African Orogen, while the stratiform deposits developed more towards the core of this belt. The stratabound Cu-Co sulphides precipitated from a higher saline (in the DRC) and mostly higher temperature fluid, relative to the vein-type ores in the front zone. Mineralising fluids in the core of the orogen locally recycled previous ores, while the mineralising brines in the front zone transported significant quantities of metals from deeper levels to the mineralisation sites, developing for example the world class Cu deposit of Tsumeb.

50 Ma after the Lufilian Orogeny developed the Kipushi Zn-Cu deposit close to the core of the Lufilian Arc from a highly saline, high temperature fluid (450 Ma in Fig. 7). This fluid is less saline and warm than the fluid from which the 2nd main Cu-Co mineralisation precipitated close to the core of the Lufilian Arc, but warmer and more saline than the fluids from which the Cu-rich vein-type deposits on the Kundelungu Plateau and the Otavi Mountain Land developed. The mineralising fluid at Kipushi acquired its high salinity by halite dissolution and its significant metal content by interaction with the underlying sedimentary pile and/or the basement. S was derived from Neoproterozoic evaporites and maybe from Neoproterozoic seawater and reduced by TSR on site, with coeval oxidation of the existing hydrocarbons. Some of this S probably mixed with diagenetic S produced by BSR and stored in the host-rock.

The Pb-Zn deposits of Kabwe, Lombe and Kengere in the Lufilian Arc and the deposit of Berg Aukas from the Otavi Mountain Land share many characteristics with the

Kipushi mineralisation. They are all situated at approximately the same stratigraphic level and hosted by massive dolomite. These deposits are, however, dominated by Pb and Zn, with minor Cu at the contact with the overlying dolomite-shale sequence (cf. Lombe). The mineralising fluid had the same high temperature at Kabwe close to the core of the Lufilian Arc, but had a lower temperature in the Berg Aukas deposit in the foreland of the Damara Belt. S in the deposit of Kabwe was not the result of TSR from Neoproterozoic seawater or evaporite sulphate, but was reduced by BSR prior to introduction of the mineralising brine. S from the sulphides at Berg Aukas was reduced through TSR, similar to the deposits of the Tsumeb-type in the same area. The source of the mineralising fluids in these deposits has been strongly debated. Several authors propose a magmatic-hydrothermal origin (Thoreau, 1928; Taylor,1954; Intiomale & Oosterbosch, 1974; Walraven & Chabu, 1994). Others believe the mineralising fluids at Kipushi and Kabwe could also be evolved connate fluids (Kortman 1972) or fluids produced by metamorphic dewatering (Unrug, 1988; Kampunzu et al., 2009). All these deposits are characterised by a low radiogenic, almost homogeneous Pb isotopic cluster in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space. The vein-type deposits of Kipushi, Lombe and Kengere in the DRC cluster at lower $^{207}Pb/^{204}Pb$ values, compared to the Berg Aukas and Kabwe deposits that have a $^{207}Pb/^{204}Pb$ isotope ratio in the range of the vein-type deposits of Tsumeb, Khusib Springs, Kombat and the Dikulushi Cu-Pb-Zn-Fe mineralisation (Fig. 6C). Pb isotope data for the deposits of Kipushi, Lombe, Kengere, Kabwe and Berg Aukas indicate Pb and perhaps other metals were derived from several sources that may include the underlying sedimentary pile and basement and their Pb was homogenised in large-scale hydrothermal systems, before injection in dolomite horizons, favourable for ore deposition. The age of formation for the vein-type deposits of Lombe, Kengere, Kabwe and Berg Aukas is uncertain. Nonetheless they are placed together with the Kipushi deposit in Fig. 7, because these deposits show many similarities.

A low to moderate saline, low temperature fluid remobilised the Cu-Pb-Zn-Fe mineralisation at Dikulushi in a Cu-Ag mineralisation (< ca. 520 Ma in Fig. 7, possibly at ca 100 Ma). This fluid is the result of mixing a moderately saline fluid that resembles the Cu-Pb-Zn-Fe mineralising fluid, with a low saline fluid that could represent meteoric water saturating the host sandstone (Haest, 2009). The Cu-Ag mineralising fluid likely remobilised metals and S from the Cu-Pb-Zn-Fe mineralisation. Addition of radiogenic Pb from carbonate or sandstone host-rocks produced Cu-Ag sulphides with a Pb isotopic composition that plots along a linear trend in the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ space. A late phase of minor pyrite-chalcopyrite-sphalerite mineralisation at Kipushi precipitated from a lower saline, lower temperature, more oxidising fluid, relative to the fluid that precipitated the main sulphide mineralisation at Kipushi. This fluid could be of meteoric origin and acquired a high salinity by

dissolution of halite. The second mineralisation phase at Kipushi is placed together with the Cu-Ag mineralisation at Dikulushi in Fig. 7, based on the possible contribution of meteoric water in the mineralising fluid.

7. Conclusion

The stratiform Cu-Co ore deposits formed during successive periods in the evolution of the Lufilian Orogen. Disseminated sulphides and sulphides cementing pores and replacing evaporites have an early diagenetic origin. This phase is related to a mineralising fluid with a moderate temperature and salinity, which most likely interacted with the basement. S originated from the bacterial reduction of Neoproterozoic seawater sulphate. Ore deposition during diagenesis in Zambia and likely in Katanga was controlled by extensional faults cross-cutting the basement and Lower Roan Group. An important stage of mineralisation/remobilisation took place during metamorphism and tectonic activity related to the Lufilian orogeny between 580 and 520 Ma (El Desouky et al., 2009, 2010). High temperature fluids were expelled within a compressional regime and mineral precipitation occurred in nodules, successive vein generations and tectonic breccia.

The vein-type Cu-Zn-Pb mineralisation in the southern and central part of Africa have many common characteristics. (1) They are hosted in massive dolomite (and sandstone), with associated organic material, at two different levels in the Neoproterozoic stratigraphy. The lower level Zn-Pb(-Cu) mineralisation are hosted by massive dolomite and these mineralisation become Cu-rich near the contact with the overlying dolomite-shale sequence. The higher level Cu(-Zn-Pb) mineralisation are hosted by often brecciated sandstone and dolomite. (2) The higher level Cu deposits that have been dated, i.e. Dikulushi and Tsumeb, are syn- to late-orogenic, while the lower level Zn-Cu mineralisation of Kipushi is post-orogenic in age. (3) All deposits are associated with steep faults that crosscut the stratigraphy, independent of their age of formation. These faults could form conduits for the mineralising fluids. (4) The source of these mineralising fluids is unknown for the deposits in the DRC and Zambia. Dewatering magma, deeply infiltrating seawater, connate water and metamorphic water have been proposed as potential sources for the mineralising fluids. The mineralising fluid for the vein-type deposits in Namibia is likely seawater. (5) Syn- to post-orogenic Cu-Zn mineralising fluids at Dikulushi, Tsumeb and Kipushi migrated upward from depth, where they acquired a high salinity from evaporated seawater or from the dissolution of evaporites and a significant metal content by leaching metals from the basement and/or from the overlying sedimentary pile. These mineralising fluids disposed their metal content, when they mixed with a reducing phase on the site of mineralisation in the form of hydrocarbons or H_2S. The deposits of Lombe, Kengere, Kabwe, Berg Aukas, Kombat and Khusib Springs probably share this history of formation. (6) The syn- to postorogenic polymetallic mineralisation are characterised by low

radiogenic Pb isotopic compositions that are the result of hydrothermal fluid circulation systems, which scavenged and moderately homogenised Pb from different host-rocks. Pb isotope data from the lower level Pb-Zn deposits are more homogeneous than for the higher level Cu deposits. The more heterogeneous Pb isotopic composition of these syn-orogenic Cu deposits could be attributed to smaller scale fluid circulation systems that did not succeed in completely homogenising the Pb isotope composition.

The mineralisation at Dikulushi and Kipushi were remobilised afterwards by oxidised, less saline, lower temperature fluids that are at least partly meteoric in origin. The Cu-Ag sulphides from the remobilisation phase at Dikulushi have higher radiogenic Pb isotopic signatures and display linear trends on both $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagrams.

8. Acknowledgements

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