Multiphase origin of the base metal deposits in the Lufilian fold-and-thrust belt,
Katanga
(Democratic Republic of Congo)

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Summary

The Central African Copperbelt hosts world class stratiform Cu(-Co) deposits and vein-type base metal deposits in the Neoproterozoic Katangan Supergroup in the Democratic Republic of Congo (DRC) and Zambia. Based on literature review, satellite image interpretations and petrographic, geochemical and fluid inclusion analyses, multiple mineralisation/remobilisation phases that occurred during different stages of the evolution of the basin have been identified for the generation of the mineral deposits.

The first and likely main period of mineralisation occurred after sedimentation in the Katangan part of the Copperbelt, during early to intermediate diagenesis of the lower part of the Roan Group. Early diagenesis started with the precipitation of framboïdal and euhedral pyrite. The main phase of the stratiform mineralisation followed pyrite precipitation and consists of disseminated copper and often cobalt sulphides and sulphides (carrolite, chalcopyrite, bornite, digenite and chalcosite) in nodules and lenses, which are often pseudomorphs after evaporites (Muchez et al. 2008). A diagenetic origin is supported by the Re-Os dating of chalcopyrite in the Konkola deposit (Zambia) at 816 ± 62 Ma (in Selley et al. 2005) and geochemical data. This mineralisation phase can be related to the early Katangan rifting of the basin during the Roan (Muchez et al. 2007), leading to the formation of a passive continental margin.

A second mineralisation and/or remobilisation phase is characterised by the occurrence of Cu and Co sulphides in dolomite and quartz veins that crosscut the nodules and lenses of the mineralisation phase. A remobilisation is suggested by the identical mineralogy of the sulphides of the first phase and in the veins and geochemical similarities between both phases. Two periods can tentatively be suggested for formation of these mineralised veins. Firstly during the Late Roan, when volcanic and magmatic rocks were emplaced in the continental rift setting and when an elevated heat flow was present(Muchez et al. 2007). Secondly during the Lufilian orogeny between 592 and 512 (Rainaud et al. 2005) that caused the deformation of the Katangan sediments. Radiometric dating of sulphides at Nkana, Chibuluma and Nchanga (Zambia) revealed mineralisation ages around 583 and 526 Ma (Barra et al. 2004). During the latter period also mineralisation of tectonic breccias in the Roan, related to the Lufilian orogeny, could have taken place. Vein-type polymetallic and Cu-Ag deposits also occur in the Copperbelt (e.g. Kipushi and Dikulushi). Structural and radiometric evidence indicate these deposits formed during and after the Lufilian orogen.

Superficial weathering of the primary sulphide deposits resulted in the formation of Cu and Co-hydroxides, -oxides, -silicates or -carbonates at the surface. In addition to an extensive set of newly formed minerals, this secondary enrichment also resulted in an upgrade of the Cu and Co content.
# Table of Content

1. Introduction 4

2. Stratigraphy of the Kundelungu Supergroup 7
   2.1 Stratigraphy 7
   2.2 Tectonic history Copperbelt 10
   2.3 Metamorphism and magmatic/volcanic rocks 11
   2.4 Temporal evolution of the Lufilian belt 13

3. Literature review 15
   3.1 Central part of the Copperbelt 15
      3.1.1 Shinkolobwe 15
      3.1.2 Anticline of Mulungwishi 16
      3.1.3 Kabolela 17
      3.1.4 Kambove west, Kamoya and Kakanda Sud 21
      3.1.5 Shituru 24
      3.1.6 Fungurume 24
   3.2 Eastern part of the Copperbelt 27
      3.2.1 Etoile 27
      3.2.2 Kipapila 31
      3.2.3 Hematite in the Lubumbashi area 34

4. Database Cu-Co mineralisation in the Katanga fold-and-thrust belt 34

5. Satellite imagery and mineralisation 36

6. Petrography of stratiform mineral deposits 39
   6.1 Western part of the Copperbelt 39
   6.2 Eastern part of the Copperbelt 42
   6.3 Central part of the Copperbelt 42

7. Geochemistry 44
   7.1 Sulphur isotope composition 44
      7.1.1 Literature 44
      7.1.2 Recent work 46
   7.2 Stable carbon and oxygen isotopes 47
      7.2.1 Literature 47
      7.2.2 Recent work 48
   7.3 Microthermometry 49
      7.3.1 Literature 49
      7.3.2 Microthermometric investigation 51
   7.4 Radiogenic isotopes 54

8. Metallogenic model for stratiform mineralisation in the Copperbelt 55
   8.1 Cu-Co mineralisation and satellite imagery 55
   8.2 Early diagenetic origin 55
   8.3 Diagenetic origin 56
   8.4 Late diagenetic and syn-to post-orogenic mineralisation 60
   8.5 Supergene enrichment 61

9. Conclusion 62

10. References 64
1. Introduction

The Katanga province is located in the southeast of Democratic Republic of Congo (DRC) and hosts part of the Central African Copper belt that is straddling the border between Congo and Zambia. This belt is the largest and highest grade sediment-hosted stratiform copper province known on earth. Combined production and reserves total approximately 190 Mt of Cu of which 102 Mt are contained in the Congolese part of the belt (i.e. the Lufilian arc), as stratiform or vein-type deposits. The stratiform mineralisation stretches from Kolwezi up to Kimpe in the DRC (Figure 1). The stratiform copper deposits have been thoroughly investigated since their recognition in the early 1900s. Most stratiform copper deposits are located in the Lufilian arc (e.g.: Luiswishi, Kamoto, etc.), except for some smaller mineralisation in the foreland (e.g.: Lufukwe, Kibodia, Mwitapile, etc., Dewaele et al. 2006, El Desouky et al. in press). In the Lufilian fold-and-thrust belt, all stratiform deposits are pre- to syn-orogenic Cu-Co mineralisations. The stratiform mineralisation has historically been mined by the Union Minière du Haute-Katanga (UMHK), later by the Gécamines.

Although the stratiform mineralisation form the largest part of the total production and reserves, the syn- to post tectonic vein-type Zn-Cu mineralisation has also played a major role in the mining history of the region. For example, the Kipushi deposit (Intiomale and Oosterbosch, 1974; Intiomale, 1983; de Magnée and François, 1987) was one of Africa’s largest producers of Zn and Cu during the previous century. It is located some 30 km to the WSW of Lubumbashi.
Since the majority of the mineralisation present in the Lufilian fold-and-thrust belt belongs to the stratiform type (see later), this type of mineralisation forms the main topic of interest for this report. This study mainly focused on the origin and genesis of the stratiform mineralisation situated in the Congolese part of the Copperbelt, between roughly Kolwezi and Lubumbashi. The Congolese part of the belt extends itself up to Kimpe. This area has historically been subdivided in a western (area of Kolwezi), central (area of Kambove) and eastern part (area of Lubumbashi).

Although numerous metallogenic studies have been carried out to deduce the origin of these deposits, their formation still remains a matter of intense debate. The models proposed include syn-sedimentary, early diagenetic, late burial diagenetic or syn-orogenic origins.

![Figure 2. Schematic representation of the syn-sedimentary/syn-genetic model (after Garlick, 1972)](image)

The syn-sedimentary or syn-genetic theory links the precipitation of sulphides to the deposition of the host-sediments. According to Fleischer et al. (1976) and Garlick (1989), the copper came from streams and reacted with hydrogen sulphide in anoxic standing water to form the mineralisation (Figure 2). The model was based on the presence of sulphidic bedding planes eroded by scour channels, slump folded ore horizons (Fleisher et al., 1976; Garlick 1962; 1989) and a zonal sequence of bornite-chalcopyrite-pyrite that follows the direction of the sub-marine currents (Garlick, 1962). However, this model has
been invalidated based on the lack of systematic correlation between transgressive/regressive events and sulphide zonation and based on the discontinuity of the mineralisation within a single lithostratigraphic unit in Zambia (Annels, 1974; Sweeney & Binda, 1994). A syn-sedimentary to very early diagenetic origin is still favoured by Okitaudji (1989; 1992; 2001) for the Cu-Co deposits of Katanga in the DRC.

Bartholomé et al. (1971; 1972) and Bartholomé (1974) proposed a diagenetic, pre-deformation origin for the Cu-Co mineralisation at Kamoto at the western end of the Lufilian fold-and-thrust belt. In this model an early sulphide generation formed during deposition and early diagenesis of the host-rock, followed by a second sulphide generation due to the interaction of the host-sediment and its pore water with a metal-bearing brine. The unknown origin of the fluid, metal and the exact timing of the mineralisation, led to numerous variations on the diagenetic model (e.g. Annels, 1989; Figure 3, Cailteux, 1974; Dejonghe and Ngoyi, 1995; Lefebvre, 1989; Unrug, 1988).

McGowan et al. (Figure 4, 2003) presented field and isotopic data for the Nchanga deposit in the Zambian sub-province that indicate an epithermal origin of this deposit, possibly during deformation of the host sequence. Wendorff (2000a) suggested that part of the stratiform mineralisation formed by the precipitation of metals in the foreland of advancing thrust sheets during the Lufilian orogeny.
The aim of this report is to demonstrate that several successive mineralisation/remobilisation phases played an important role in the formation of the economic Cu-Co ore deposits in the Copperbelt in Katanga, and not a single one. It is based on a compilation of available data from unpublished archives, combined with an interpretation of satellite imagery, the reconstruction of the paragenesis of different deposits, geochemical data and a microthermometric study of fluid inclusions in authigenic quartz, associated with the main mineralising phase.

2. Stratigraphy of the Kundelungu Supergroup

2.1 Stratigraphy
The sediments in the Lufilian belt and its foreland were deposited during the Neoproterozoic, on a Palaeo- to Mesoproterozoic basement. The Katangan sediments started to be deposited in an intra-cratonic rift (Porada and Berhorst, 2000; Unrug, 1988) or in an epicontinental marine embayment (Jackson et al., 2003). The underlying pre-Katangan basement is poorly studied in Katanga and what is known in northern Zambia has been documented by Key et al. (2001) and Rainaud et al. (2002). These sediments were deformed during the Lufilian orogeny (ca 560 – 550 Ma; Cahen et al., 1984; Kampunzu and Cailteux, 1999; Porada and Berhorst, 2000). The Lufilian fold-and-thrust belt and its foreland are bordered to the west by the Mesoproterozoic Kibaran belt and to the east by the Paleoproterozoic Bangweulu block.
The Katanga Supergroup consists of a 5 to 10 km-thick sequence that can be subdivided into three groups based on two regionally extensive diamictites (Figure 5). From the bottom to the top, the Katanga Supergroup is divided into the Roan, the Nguba and the Kundelungu Groups (Cailteux et al., 2005). Sedimentation of the Katanga system started in a continental Roan rift basin after ~880Ma (Armstrong et al., 2005), with a basal conglomerate (Cailteux, 1994). The Roan is divided into four subgroups, i.e. the R1 to R4 Subgroups. The R1 Subgroup, known as the “roches argilo-talqueuses (R.A.T.)”, consists essentially of massive or irregularly stratified detrital formations with hematite present as authigenic plates and red pigment, attesting to the primary oxidising conditions (Cailteux, 1994).

<table>
<thead>
<tr>
<th>KAROO AND KALAHARI</th>
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<tr>
<td>+/- 560-550 Ma</td>
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<tr>
<td>Group</td>
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<tr>
<td>Plateaux (Ku-3)</td>
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<tr>
<td>Kintou (Ku-2)</td>
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<td>Kalule (Ku-1)</td>
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<td>Monwezi (Ng-2)</td>
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<td>Likasi (Ng-1)</td>
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<tr>
<td>Mwashiya (R-4)</td>
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<td>Dipeta (R-3)</td>
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<td>Mines (R-2)</td>
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<td>R.A.T (R-1)</td>
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<tr>
<td>Lithologies</td>
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<tr>
<td>Shales and arkoses</td>
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<tr>
<td>Dolomitic shales, sandy shales and sandstones</td>
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<td>Dolomitic shales or sandy shales, pink limestones, Diamictite</td>
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<td>Dolomitic or sandy shales; dolostones or limestones Diamictite</td>
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<tr>
<td>Dolomitic shales Dolostone, jaspers and pyroclastites</td>
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<tr>
<td>Interbedded dolostones argillaceous and dolomitic siltstones</td>
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<td>Dolostones; dolomitic shales and siltstones</td>
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<tr>
<td>Argillaceous dolomitic siltstones; sandstones and pelites</td>
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Figure 5. Stratigraphy of the Katanga Supergroup in Democratic Republic of Congo (modified after Cailteux et al., 2005).

Towards the contact with the R2 (Mines Subgroup), there are indications that sedimentation took place in an evaporitic environment. The sedimentary transition to the Mines subgroup is at certain localities continuous, whereas at other localities a tectonic breccia developed at the contact. The tectonic breccia formed during detachment of the Mines Subgroup, which was aided by fluidisation of evaporitic material, probably present...
near the top of the R.A.T. subgroup (Cailteux and Kampunzu, 1995). The importance of salt tectonics to explain the large observed breccia bodies in the Lufilian belt is further stressed by Jackson et al. (2003). The Mines Subgroup (R2) is a carbonate unit that contains the richest stratiform copper-cobalt mineralisation, which occur at two different stratigraphic levels. The overlying R3 (Dipeta Subgroup) is subdivided in four formations, each characterised by predominantly argillaceous and siliciclastic beds at the base and by predominantly carbonate beds at the top (Cailteux, 1994). The transition to the overlying Mwashya Subgroup is again marked by a tectonic breccia which developed at the contact. The Mwashya was deposited between 760 and 735Ma (Master et al., 2005). The continuous stratigraphic sequence for the Roan group, as proposed by François (1974) and Cailteux (1994), is contested by some authors. Porada and Berhorst (2000) agree that the R.A.T. was deposited on the pre-Katangan basement, but they believe that the overlying Mines, Dipeta and Mwashya subgroups form a platform facies association, which became a Copperbelt-type tectonostratigraphical succession through the development of foreland propagating thrust faults. Wendorff (2000a+b, 2005) does not interpret the brecciated contacts as of tectonic origin. He proposed they should be called conglomerates that were derived from erosion of advancing thrust sheets during the Lufilian orogeny. The R.A.T. and the Dipeta Groups are olistostromes, deposited at the border of the developing Lufilian orogen (Wendorff, 2002a+b, 2005).

In contrast to the disagreement about the stratigraphy of the Roan, all authors agree that sedimentation of the Nguba group started with the deposition of a diamicite (the ‘Grand Conglomérat’) that likely formed part of the Sturtian glacial deposits (Kampunzu et al., 2005) (Table 1). The ‘Grand Conglomérat’ thickens towards the north (François, 1974) and is at least observed as far north as Pweto, which is a town to the north of Dikulushi (Cahen, 1954). The Likasi Subgroup, with at its base the ‘Grand Conglomérat’, contains a mixture of shale and dolomite in the south and a lateral facies change towards pure shale in the north of the Kundelungu basin (François, 1974; Batumiké et al., 2006). The Kakontwe limestone, in the middle of the Likasi Subgroup, is only observed in the south, towards the Lufilian belt. This limestone formation could be a cap carbonate, confirming the interpretation of the ‘Grand Conglomérat’ as a glacial tillite (Porada and Berhorst, 2000). The Monwezi Subgroup was deposited towards the end of the Nguba and consists of more detrital lithologies, with relatively thin arkose layers in the north and thick, slightly carbonatised pelites and fine sands in the south (François, 1974; table 1). Transition to the overlying Kundelungu Group is again marked by a diamicite (the ‘Petit Conglomérat’) that forms part of the Kalule Subgroup. This conglomerate layer could have been deposited during the Marinoan-Varanger ice age. The Kalule Subgroup is similar to the Likasi Subgroup, since it also contains a thin cap carbonate (the ‘Calcaire Rose’) overlying the ‘Petit Conglomérat’. The Kalule Subgroup becomes more detrital towards its top, with the deposition of dolomitic siltstones, sandy shales and pink oolitic limestone. The ‘calcaire rose’ is, unlike the Kakontwe limestone from the Likasi Subgroup, continuous from the south to the north, with a layer thickness of around 5m (François, 1974). The Kiubo Subgroup overlies the Kalule Subgroup and consists of sandstones and shales (Cailteux et al., 2005). The top of the Kundelungu group is formed by the Plateau Subgroup that consists of shales and arkoses.
Historically, the mineralized sections of the R.A.T. and the Mines Subgroup have been subdivided in different sections based on their appearance (e.g. François, 1974). The R1 Sub-group forms the so-called R.A.T. rouge. The Mines Subgroup (R2) has been subdivided into three formations that are subdivided into different niveaux. Formation R2.1 is subdivided into niveau R2.1.1 or the so-called R.A.T.gris, in niveau R2.1.2 that consists of subniveau R2.1.2.1 or DStrat and subniveau R2.1.2.2 or R.S.F. and niveau R2.1.3 or the R.S.C. The formations R2.2 and R2.3 are also called S.D. and C.M.N, respectively.

2.2 Tectonic history Copperbelt

Folding and overthrusting, related to the Lufilian orogeny, developed a mountain chain in the south of Katanga and some smooth anticlines (i.e.: Lufukwe-, Kiaka-, Kabangu-anticlines) in its foreland (Haest et al. 2007). Unrug (1983) divided the Lufilian belt in 5 structural domains (Figure 6), i.e. the external fold-and-thrust belt (I), the "Domes area" (II), the "Synclinalor belt" (III), the "Katangan High" (IV), and the "Katangan Aulacogen" (V). The latter forms a triangular-shaped, largely tabular succession of sedimentary rocks, and can be regarded as the orogenic foreland. Areas I and II in Figure
6 make up the Lufilian Arc sensu strictu (Kampunzu and Cailteux, 1999). The areas are considered to be composed of a stack of thin-skinned thrust sheets transported on a detachment plane that cuts the basal part of the Roan. Upright or outward verging tight folds with axes traceable for distances of 50 to 175 km are typical within the thrust sheets. The thrust sheets occur together with large megabreccia that may have a tectonic origin (Lefebvre, 1980; François, 1974; Cailteux, 1990; Cailteux and Kampunzu, 1995; Jackson et al., 2003) or a sedimentary origin for at least some of the breccia (Wendorff, 2000a; 2005).

The Katangan basin was closed and deformed during the Lufilian orogeny, leading to development of predominantly north-verging folds, thrusts and nappes. In a revision of the tectonic evolution of the Lufilian Arc, Kampunzu and Cailteux (1999) recognized three distinct phases of deformation. A first phase (D1, or "Kolwezian phase") produced regional folding, thrusting and northward directed nappe transport. Kampunzu and Cailteux (1999) suggested that this phase peaked around 790 to 750 Ma, but recent U-Pb zircon dating on volcanic rocks interstratified with the "Grand Conglomérat" indicates that these lithologies, which are deformed during the D1 event, are on average ~735 Ma (Key et al., 2001). As a consequence, the D1 event must be placed significantly later. The second phase (D2, or "Monwezian phase") was dominated by strike-slip deformation associated with sinistral movement along the ~EW-oriented Monwezi fault zone, and ultimately along the regional WSW-ENE trending Mwembeshi fault zone, an intraplate shear-zone (Figure 6). It was suggested that this caused left-lateral mass extrusion of the Katangan cover rocks. It resulted in renewed folding and faulting with production of faults that often cut parallel across D1 fold hinges ("failles d'extrusion" of François, 1974, 1987). D2 strike-slip deformation gave rise to the general convex geometry of the Lufilian Arc. Syntectonic igneous rocks along the Mwembeshi Fault Zone (Hook Granite massif) were dated by Hanson et al. (1993) at ~ 550 Ma. A third deformation phase (D3, or "Chilatembo phase") finally caused late, transverse folding and NS or EW trending faults. Many authors have stressed the importance of regional-scale salt tectonics, involving partly presumed salt-bearing Roan lithologies, in the kinematics of the Lufilian fold-and-thrust belt (e.g. François, 1974, 1987; Cailteux and Kampunzu, 1995; Kampunzu and Cailteux, 1999; Jackson et al., 2003).

2.3 Metamorphism and magmatic/volcanic rocks
The Domes area in Zambia obtained an eclogite facies metamorphism and high-pressure talc-kyanite whiteschist metamorphism (zone II on Figure 6). This metamorphism has been dated at ~590 Ma and ~530 Ma (Rainaud et al., 2005). A later regional metamorphic phase was recognised at ~512 Ma (Richards et al., 1988; Torrealday et al., 2000; Rainaud et al., 2005). The latter phase could be related to hydrothermal activity. In the Lufilian fold-and-thrust belt (zone I on Figure 6), regional metamorphism grades from prehnite-pumpellyite facies in the outer zones to lower greenschist facies towards the Domes area (Lefebre and Patterson, 1982; Cluzel, 1986). However, the metamorphism can be locally disturbed due to hydrothermal activity. Studies of the metamorphism of Roan lithologies showed indications for a sodic (paragonite + albite), potassic (muscovite + phlogopite), magnesian (clinochlore + dolomite + magnesite), and chlorine (scapolite) alteration (Cluzel, 1986; Vaes, 1962).
Numerous dolerite sills have been identified in the Dipeta Sub-group. Also spilitic lavas and interstratified tuffs are found in this stratigraphic horizon (Lefebvre, 1973). They have been submitted to a retrograde metamorphism and hydrothermal alteration. Some of the sills are associated with a Cu-Co mineralisation. However, no direct relationship is supposed/observed with the mineralisation present in the Series des Mines.

- Mwadingusha: In the upper part of the Roan/Dipeta, there occur green-coloured crystalline rocks. They occur in the axe of the anticlinal structure. Microscopic investigation indicates the presence of amphibole and plagioclase. Ilmenite, pyrite and small grains of chalcopyrite are found. Lefebvre (1956) identified these rocks as gabbros. Spectrometric analysis indicates the presence of copper and titanium and a little cobalt and silver.

-Kakonge is situated at the western side of the route that connects Luambo and Bunkeya. It hosts two spots with a Cu-Co mineralisation. It is situated in the central part of an anticlinal structure interpreted to consist of Roan rocks. The deposit of Kakonge-Pindi is the most important mineralisation. The igneous rocks, themselves can contain local enrichments of malachite and cobalt sulphates. The highest grades are found at Kakonge-Pindi in sedimentary rocks in contact with the mafic rocks. The sedimentary rocks belong to the Dipeta, on which the intrusion of mafic rocks only resulted in a very weak contact aureole. The mafic rocks are very similar to those observed at Mwadingusha. They consist of plagioclases, biotites and amphiboles. The ilmenite shows exsolution laminae from hematite and rutile. If pyrite is absent, chalcopyrite and bornite are present throughout the entire rocks (maximum size of 10µm). These latter minerals are cemented by chalcocite, digenite and covellite.

-Sambwa: The mineralisation at Sambwa occurs in an anticlinal structure that consists of Upper Mwashya rocks surrounded by Grand conglomerate. The extrusions are filled with a breccia and a large part of Dipeta rocks that consist of violet sandstones and pelites with hematite and some dolomite lenses. An oxidised mineralisation with high grades but low extension has been found. The original magmatic rocks are rarely found, since they have been brecciated. The original rocks have been obliterated by later hydrothermal activity. It is south of Sambwa toward Ruashi, that the first appearance of biotite occurs.

- In the central part of the Roan extrusion at Shinkolobwe, igneous rocks have been identified (signal Lalumbi). They seem to be related to CMN/Dipeta rocks, but there is a brecciated horizon obliterating the direct lithostratigraphic contact. It is interpreted that there is an intrusion of mafic rocks in the upper part of the CMN (similar as at Tenke and Fungurume). The contact between the mafic rocks and the host-rock is non-transgressive. The mafic rocks are crosscut by numerous quartz veinlets that contain chalcopyrite mineralisation.
- The hills of Makawe belong to the last anticlinal structure that is still visible in the Katanga Copperbelt. There is a copper mineralisation in the yellow dolomites of the lower Mwashya. The mineralisation consists of chalcocite and malachite that occurs along joints. The core of the anticlinal structure contains brecciated rocks. Mafic rocks are found in the northern part of the anticlinal structure and in the axial breccia. The rocks are dolerites. There is a low Cu grade, mostly in spots.

- Kipushi: at level 240, there is an intersection of the Kakontwe limestone and schisto-dolomitic breccia that crosscut mafic rocks. No contact metamorphism is present between the mafic rocks and the breccia. The mafic rocks are classified as a dolerite. It seems that the mafic rock was originally emplaced in Dipeta rocks.

- Diorite in the Kolwezi nappe. At Dikuluwe, a brecciated dioritic rock has been observed. The rock is hosted in brecciated rocks of the R.A.T. The dioritic rock is strongly altered and partly brecciated. No contact metamorphism is observed at the contact of the dioritic fragments. According to one hypothesis, it is a fragment detached from the basement. According to another model, it is an intrusion in the R.A.T. rocks. If the latter case, it would be an exception for the whole Katanga.

Other locations where possible mafic intrusions have been described: Kipese, Kampesimpesi, Muombe, Luishia, Midiashi, Kavundi west

2.4 Temporal evolution of the Lufilian belt
Fundamental to understand the relationship between the geodynamics of a basin and ore deposits is the absolute dating of the mineralising periods (see also Muchez et al. 2007). The formation occurred definitely after ~880 Ma (877 ± 11 Ma, Armstrong et al. 2005), since this is the crystallization age of the Nchanga granite, on which the sedimentation of the Katanga system started. Although it is often difficult to date ore deposits in sedimentary rocks, few mineralisation in the Central African Copperbelt have been well-dated. Numerous U-Pb and K-Ar age of uraninite mineralisation in the Congolese Copperbelt are available in the older literature (Cahen et al., 1961; Cahen et al., 1971; Cahen et al, 1984). But these systems are susceptible for resetting. Therefore, these ages should be carefully interpreted. Only recently, techniques that are more resistant to resetting have been used for the dating of mineralisation in the Copperbelt. Preliminary Re-Os isotope dating on the Konkola stratiform Cu deposit suggests a mineralisation at 816 ± 62 Ma (in Selley et al. 2005). This age confirms an episode of diagenetic mineralisation in the Copperbelt. A five point Re-Os isochron with analyses of sulphides from the Nkanka, Chibuluma and Nchanga ore deposits yields an age of 583 ± 24 Ma. This age overlaps with the oldest biotite (~586 Ma) and monazite (~592 Ma) ages and with the synorogenic magmatic rocks (~560 Ma). Also a younger Re-Os molybdenite age of 525.7 ± 3.4 Ma has been reported from Nkana (Barra et al., 2004). This age is comparable to the 531 Ma of the high-pressure talc-kyanite whiteschist metamorphism (Rainaud et al., 2005).
Late hydrothermal alteration and veining at the Musoshi stratiform copper deposit in the Democratic Republic of Congo has been dated by U-Pb rutile and uraninite geochronology (Richards et al., 1988a, b). Six rutile samples gave a U-Pb concordia upper intercept of 514 ± 2 Ma, which was confirmed by nine analyses of five uraninite vein samples that showed an age of 514 ± 3 Ma. The fluid inclusions in the vein quartz from the hydrothermal zones consist of a halite-saturated liquid at ~400°C with approximately 39 wt% NaCl and 15 wt% KCl and a minor amount of CO$_2$ and other components (Richards et al., 1988a). This hydrothermal activity is interpreted to be related to the compressional deformation and metamorphism associated with the Lufilian orogeny.

U/Pb dating of two uraninite generations at Luiswishi by Loris et al. (1997) provided an age of 625 ± 5 Ma and 530 ± 0.9 Ma. The former age is situated at the transition between the Nguba and the Kundelungu and the latter lies in the range of the Lufilian orogeny in the DRC.

The Kansanshi copper deposit in northern Zambia consists of high-angle, sheeted quartz-carbonate-sulphide veins with envelopes of disseminated sulphides (Torrealday et al. (2000). The veins cut and replace metamorphosed Katangan sedimentary rocks of Neoproterozoic age. Based on crosscutting relationships, three stages of subparallel veins were identified. The first two vein sets are rich in chalcopyrite and contain minor molybdenite. The third vein set is characterised by relatively abundant molybdenite with significant monazite and brannerite and minor chalcopryite. Re-Os dating of the latter molybdenite-monazite vein set indicates an age of 502.4 ± 1.2 Ma, 10 million years younger than the dominant chalcopyrite veins (512.4 ± 1.2 Ma; Torrealday et al. 2000). SHRIMP U-Pb analyses of monazite from the third vein set yield a U-Pb age of 511 ± 11 Ma, within error similar to the Re-Os age. The ages of the mineralisation at Kansanshi are broadly similar to those determined for other post-tectonic vein systems in the Central African Copperbelt. A major mineralisation event occurred during and after peak metamorphism and mineralising fluids responsible for the formation of many of these deposits may have been metamorphic in origin (Greyleng et al. 2005).

An age of ~800 Ma for the stratiform Cu mineralisation in the Central African Copperbelt clearly indicates a Roan age for the mineralisation (the error on the age determination of ~60Ma could still indicate a Nguba age). Petrographic arguments clearly indicate mineralisation started shortly after sedimentation and continued during shallow burial (Muchez & Dewaele, 2006).

An important period of magmatic activity, characterised by volcanic and pyroclastic rocks, occurred during the Dipeta (Cailteux et al., 2007). Differential subsidence took place during this period (François, 1973), possibly initiated along syn-sedimentary faults (cfr. Porada & Berhorst, 2000). Such extensional tectonic regime with high heat fluxes is an ideal setting for sediment-hosted Zn-Pb and Cu deposits, including sedimentary exhalative, diagenetic and epigenetic type deposits (Muchez et al., 2005). The available radiogenic data from both the magmatic activity and the Cu mineralisation indicate that the first main phase of mineralisation largely pre-dated this magmatic activity (760 Ma). However, important thickness variations in the Lower Roan rocks are present (François,
1974), reflecting differential subsidence started earlier during the Roan. In addition, the second mineralisation phase characterised by the occurrence of coarse-grained dolomite and sulphide minerals in veins and their surrounding host-rock could have taken place during this period of magmatic activity.

Unrug (1988) proposed mineral emplacement during the Nguba by convective circulation of basinal brines forced by a thermal gradient created by a wedge-shaped lithosome of Nguba pelitic rocks with low thermal conductivity. Igneous intrusions and volcanic rocks were interpreted as the source of metals. The latter view is largely questioned by Cailteux et al. (2005) who stressed that mafic rocks are much less abundant than the 50% stated by Unrug (1988).

Petrographic and structural evidence and radiometric dating all clearly indicate that a main stage of ore deposition between 580 and 510 Ma (Cu-Zn-Pb-Ag-U) occurred during metamorphism and tectonic activity related to the Lufilian orogeny. Recent studies in the Lufilian foreland also indicate that mineralisation continued after this orogeny (Dewaele et al., 2006; El Desouky et al., 2006; Haest et al., 2006).

3. Literature review

In addition to literature that describes the more general setting (stratigraphy – tectonic history – metallogenic model), some mineral deposits - situated between Kolwezi and Lubumbashi - have been studied in more detail. These studies were, however, mostly oriented towards a characterisation of the local stratigraphy and/or the mineralogy of the deposits. In addition, superficial descriptions of mineral deposits in the western and central part of the Congolese Copperbelt can be found in François (1973) and François (2006).

3.1 Central part of the Copperbelt

3.1.1 Shinkolobwe
The deposit of Shinkolobwe is characterised by the presence of a suite of metals: Ni, Co, Cu, Fe, V, Mo, Au, Pt, Pd and Se occur associated with the uranium deposit of Shinkolobwe. Numerous detailed studies have been carried out in the past to characterise the complicated mineralogy. Deliens (1975) made detailed descriptions of different rare uranium minerals, like uranium molybdates (iriginite, umohoite, a Mg-Ca umohoite and zippeite) that occur as fine-grained crusts or as tine plates on massive uraninite and molybdenite that is associated with siegenite, wulfenite and quartz.

Similar mineralisation has also been described at Swambo and Kalongwe. (Derricks et al., 1958). Uranium has been found at more than 20 places in the southern part of Katanga (François, 1974), e.g., Kolwezi, Musonoi, Kamoto, Mashamba, Kalumbwe-Muunga, Mashitu, Kambove and Luishya. Uranium mineralisation occurs along the southern margin of the Copperbelt, from Kalongwe in the west to Lukuni in the east, by passing through Menda, Kasompi, Swambo, Shinkolowbe and Luishya. The uranium mineralisation is hosted in the Series des Mines, the highest concentration is found in the
DStrat and the RATgris and at the margins of the RSC. Sometimes the uranium is present as uraninite crystals, but can most often only be identified by radiometric measurements. The value is about 10 to 40 times the background value (~ 20µR/h). The uraninite is disseminated in the host-rock and in fissures and cracks. Mostly nickel (siegenite, vaesite, catvierite) and iron are associated with the uranium minerals. The alteration associated with the uranium mineralisation can reach a depth between 100m and 500m. The intensity of the alteration is more pronounced along fractures than in the bulk of fragments of Roan rocks. At the hydrostatic surface, the highest concentrations of uranium are found.

Ngongo-Kashisha (1975) concluded that the interaction between tectonic deformation followed by supergene weathering resulted in the uranium deposits. According to Derricks and Vaes (1956), the uranium mineralisation is due to ascending mineralising fluids related to magmatic source. The mineralisation was emplaced along fractures in the Roan rocks.

3.1.2 Anticline of Mulungwishi
The anticlinal structure of Mulungwishi is situated 25 km to the northwest of Likasi. In the anticlinal structure, the northern limb has been thrusted above the southern limb. A borehole has been emplaced in the central part of the Mulungwishi anticlinal structure (Lefebvre 1975). Some of the horizons found are poorly known or have never previously been described. They include spilitc pyroclastic rocks, ignimbrites and a polygenetic conglomerate. The borehole was implanted to cut the galena and chalcopyrite mineralisation that is present in the upper part of the Roan (Mwashya) and the malachite in the axial part. The borehole gives a continuous section from the Grand conglomerat to the Mwashya (“Groupe de Kansuki”).

The rocks consist mainly of a transition from white dolomite with cherty nodules at the base, passing into dolomites with stromatolites, sandstones, pyroclastic rocks, conglomerate, (carbonate-rich) pelites and dolomitic limestones. A detailed stratigraphic description of the rocks is given below:

-Silica-rich dolomites; dark dolomites with several siltite layers. Presence of nodules.

-Conglomeratic levels;
  * pyroclastic formation: change of colour. This rock consists of quartz, microcline and plagioclase (little biotite, white mica and chlorite). There has been a silicification of the rocks. Pyrite is present in these rocks. Copper is also present as small inclusions in the lapilli. The copper minerals are more present in the quartzo-dolomitic veins (chalcopyrite, bornite, digenite and chalcocite).
  * polygenetic “poudingue”: this is a micro-conglomerate, with varying composition: dolomite, microcline and quartz. This sequence has been interpreted as similar to the typical “Pisolites noires” of the Mwashya
-“Sequence negative inférieur” consists of a fine-grained dolomite and siltite with varying darker and brighter colours. Pyrite is abundant, with often chalcopyrite. The dolomitic levels are enriched in debris of volcanic rocks. The pelitic layers are enriched in mica, chlorite and rutile. This sequence passes to a siltite that is very similar to the SD of the Series des Mines. Also pyrite and chalcopyrite are found in these sediments.

-“Sequence negative supérieur”. This sequence is similar as the previous but is more darker due to the disappearance of the lighter-coloured beds.

-“Pelites dolomitiques supérieurs” is situated in the meteoric altered zone. This package is very homogeneous with only fine quartzite layers in the upper part. These quartzites are considered to be similar to those observed in the upper part of the Mwashya in the east and southeast.

3.1.3 Kabolela
The Kabolela mineralisation is situated in the central part of the Copperbelt. It contains Cu-Co mineralisation, but locally also Zn. The mineralisation is interpreted to be due to an epigenetic remobilisation on top of a weak copper mineralisation of diagenetic origin (Lefebvre 1975). It has been interpreted to have an early deformation age (Mwashya) and to be related to the intrusion of the mafic sills that are described to be mineralised in copper/cobalt and that intruded the Dipeta rocks. Kabolela is situated in the Menda facies (François, 1975), which is normally poorly mineralised, but contains locally interesting spots of copper in the upper orebody and copper-cobalt in the lower orebody. Mineralised fragments of the Series des Mines are present, together with unmineralised fragments in a mixture of Roan rocks (Figure 7).

Stratigraphy
The stratigraphy of the sediments of the Series des Mines is very similar as in Kolwezi or Lubumbashi, but there are strong variations in thickness. Even visible from borehole to borehole.

- The CMN is rarely cut by the boreholes and is badly known. The morphology is similar as the rocks lower in the sequence, but massive dolomite are more dominant. The thickness of the CMN is strongly variable. The transition to the SDS is placed at the contact between dolomitic rocks and sandy mica-rich, dolomitic pelites. The boundary is characterised by the presence of organic material. The margin is imprecise. SD3b has a varying thickness and is not always present.

- The subdivisions of the different members of the SD is not so clear.

- The RSC of Kabolela is remarkable lenticular. In the thickest part, the stromatolitic textures are perfectly visible in a quartzo-dolomitic ensemble. In the centre of these organic massifs, there are locally organic-rich pelites.
The RSF consists of dolomite and quartz and has a banded texture. The rocks are dark grey. Detrital mica is often present. At the base, a dark horizon with silica and oolites, similar as in Etoile, occurs.

Due to the absence of the black siliceous layer at the base of the RSF it is difficult to determine the boundary of the DStrat. Only the thickening of the different bands and the appearance of large cherty nodules makes it able to identify this facies. A sedimentary breccia has been described at the base. Recrystallised pyrite has been described towards a zone of deformation. This pyrite contains inclusions of chalcopyrite and bornite.

RAT gris consists of massif grey/green-grey rocks, with more chlorite-rich and silty parts. Zones with recrystallised dolomite, quartz, leuchtenbergite and authigenic mica are described.
Mineralisation

-Kabolela Nord (Figure 8) is considered as barren with grades below 2%. Lefebvre describes a stratiform pyrite-rich mineralisation with traces of copper (0.2% of Cu). A copper-rich mineralisation that consists of vein-type mineralisation of chalcopyrite associated with dolomite and dahlite. The little amount of cobalt occurs as fine bands of carrolite around chalcocite.

- Kabolela Est (Figure 8) shows a similar mineralisation type that occurs in small spots. There is mainly pyrite, with small inclusions of chalcopyrite and bornite. The ensemble is surrounded by chalcopyrite and in the alteration zones spots of carrolite and iron oxides occur. In the upper part of the RSC (borehole 130), there is a strong enrichment of Cu and Co. It consists of an association of chalcopyrite, bornite, chalcocite and carrolite that seems to replace coarse pyrite. In borehole 131, the copper mineralisation is present as chalcopyrite in quartz/dolomite veins with dahlite. Chalcopyrite is also present in the upper part of the R.A.T., DStrat, RSF and RSC, each time as an aureole around pyrite grains.

- Kabolela Sud (Figure 8) is characterised by the presence of a mineralised lens in the RSC, with little mineralisation in the RSF and the DStrat. The mineralisation is massif in a completely recrystallised gangue. The grade can be up to 4.5 to 10%. Zinc is present is present in a coarse brecciated rock.

- The main deposit (gisement principal). In the northern margin, the mineralisation is weak (max 1%). The mineralisation consists of pyrite, with inclusions of chalcopyrite and bornite, of chalcopyrite as an overgrowth on pyrite and in dolomite/dahlite veins. Pyrrhotite and millerite have been described in borehole Kab134 (Figure 8). Cobalt sulphides surround chalcocyprire. In the Kab 114 borehole, the rocks are cut by numerous dolomitic veins with pink siderite and idiomorphic pyrite, with traces of chalcopyrite. It seems that the veins with quartz, dahlite, chalcopyrite, bornite and carrolite are crosset by veinlets with chalcopyrite and siderite.

- In borehole Kab 133, the mineralisation is present as veins in the RAT gris. The mineralisation is present in veins that are associated with chlorite. The mineralisation consists mainly of chalcopyrite and pyrite. Other traces of simple Cu-sulphides and carrolite with inclusions of pyrite have been found. The pyrite is disseminated throughout the rocks, while the Cu-Co sulphides are present in veins. Sometimes, the pyrite contains inclusions of bornite.
Different phases are identified in the mineralisation.

1. Primary pyrite with inclusions of chalcopyrite and bornite. Found in the entire Series des Mines, except RSC
2. Idiomorphic pyrite found in quartzo-dolomitic recrystallisations, often with bornite inclusions.
3. Sphalerite in quartz-dolomite veins in RAT gris and in RSC. Bornite is posterior to pyrite, but is replaced by a new bornite generation and chalcocite.
4. Chalcopyrite in dolomite veins with dahlite. These veins cut the entire Series des Mines. The chalcopyrite impregnates the host-rock and occurs also as overgrowths on existing pyrite. Millerite and pyrrhotite are found with this phase.
5. Carrolite is superposed on chalcopyrite and pyrite. There is also a change from bornite to digenite.
6. This phase is considered as a crystallisation of covellite, associated with digenite and chalcocite. These minerals occur as rims of existing mineralisation. An alteration due to meteoric waters has been proposed for this phase.

Fracturating of the rocks seems to have occurred in two steps. In a first step, the main fracturing occurred at the top of the CMN and at the base of the RAT gris. This isolated the middle part of the Roan, which made it more accessible for mineralising fluids. In a second step, a large brecciation happened of the rocks below RAT gris and above the CMN. Demesmaecker et al. (1962) place the mineralisation prior to the main Lufilian deformation. The mineralisation is linked to structures that formed prior to this main event, a Mwashya age is proposed (angular discordance at the top of the Kitondwe groupe (Dumont 1971), local erosion of the Roan of which the fragments have been incorporated into the Ng1.1 (Cahen, 1974). There is a description of Cu-Co mineralisation associated with the volcanism of the lower Mwashya (Lefebvre, 1974), linked to the mafic sills that are intruded into the Dipeta. François (1974) speaks of an epithermal mineralisation related to a “hypovolcanism” contemporaneous with the deposition of the Mwashya, superposed on a local syn-diagenetic/volcano-sedimentary precursor.

3.1.4 Kambove west, Kamoya and Kakanda Sud
The stratigraphy of the Kambove west, Kamoya and Kakanda Sud deposits (Figure 9) has been studied in detail by Cailteux (1997). The transition from RAT red to RAT gris consists at these locations of a mottled rock of identical composition. The colour is just due to the change in redox conditions. The top of the R1 and R2 consists of pyroclastic rocks that were deposited in a carbonate environment. Similar situations exist in DStrat, RSF, RSC, SDB, SD2a and sporadically in SDS. The breccia below the RAT red consists of fragments going from RAT to Ku2. These fragments are gathered within a dolomite matrix. The lower part of the RAT red is often fractured. The RAT is red coloured due to the pigmentation of hematite. This hematite colours diagenetic minerals, such as authigenic quartz, dolomite and growth zones of tourmaline. The RAT gris is characterised by the fine dispersion of fine crystalline pyrite. The DStrat consists of a well stratified silicified dolomite, with nodules. The DStrat has an average thickness of 2 to 3m. The nodules consist mainly of a quartz/chert core surrounded by a carbonate rim. If the diameter is less developed only carbonate is present. In Kambove west, Kamoya 10 and Kakanda Sud this dolomite consists mainly of dark crystals. Sometimes magnesite is reported. The presence of uranium is observed with this magnesite. The RSF consists of a
rapid succession of fine-grained dolomitic, quartzo-cherty, quartzo-dolomitic and chloritic-quartz beds.

Figure 9. Description of the stratigraphy in the Kambove area (Cailteux, 1997)
The rocks contain numerous nodules that consist mainly of quartz with a rim of sulphides, or of quartz-dolomite. Cailteux (1997) identifies a second phase of silicification that overgrows the sulphides. The RSC consists of a silica-rich massive dolomite with well visible stromatolitic structures. The RSC is not homogeneous. There are numerous intercalations of more chlorite-quartzitic layers. It seems that the mineralisation is preferentially concentrated in these fragments. The SD forms an ensemble of fine-grained dolomite, but also with a detrital part (quartz, muscovite and tourmaline). There are recrystallisations of quartz that contains dolomite inclusions. There are also dolomitic and quartzo-dolomitic nodules. The SDB is typically characterised by the presence of these nodules that can be up to 1cm large. The composition is mostly quartz, sometimes quartz/dolomite. Most of the rocks are characterised by a slight fracturing.

The origin of the rocks of the RAT gris has been determined, mostly based on the authigenic Mg-rich chlorite. Part of the rocks came from the Kibara and consists of quartz, green and bleu tourmaline, green to brown chlorite, ilmenite, biotite and illite. Different theories exist about the origin of the chlorite:

1. recrystallisation of palygorskite that formed in Mg-rich lakes, associated with dolomite, sepiolite, chert, aragonite, chert, hydromagnesite.
2. authigenic chlorite formed by the transition of detrital illite (Katekesha, 1945).
3. deposition of volcanic ashes in a very reactive place; see Etoile;

Primary dolomite is present. Inclusions of calcite have been described in dahlite, so it must have existed.

The CMN is the upper part of the R2.3. It divided in a lower part (CMN inférieur R2.3.1) that is darker coloured due to the presence of organic material. The upper part (CMN supérieur, R2.3.2) is brighter and has intercalations of dolomite- and chlorite-rich sandstones.

The lower CMN is subdivided into three parts:
- unit 1: dolomite with algal structures. Nodular structures occur.
- unit 2: finely banded dolomite, often a bit organic at the bottom. Above the first 2 to 5m, the rocks have a tigre print. The bands consist of chert and coarse-grained quartz that contain inclusions of dolomite or dolomite with quartz inclusions that are parallel to the bedding. Quartz with an undulose extinction is present;
- unit 3 consists of badly stratified dolomite that shows a kind of boudinage, augen texture. Locally rock with a syn-sedimentary breccia texture appear, also slumps are visible;

The upper part of the CMN also consist of three parts:
- unit 1 consists of a white to pink brown badly stratified dolomite
- unit 2 consist of stratified dolomite, with numerous bancs of stromatolites. Almost all the rocks that form the CMN inferior can be found back. Also clearly detrital bancs can be found back;
- unit 3 consists of polychrome dolomite.

A section of grey-coloured rocks are identified as Dipeta. They are dolomitic and phyllitic. Authigenic quartz is present but is only a minor constituent (<5%).
In general, the sedimentology and the stratigraphy of the R2.3 is very similar in the central part of the Copperbelt (Kamoya, Kakanda, Mirungwe, Kamatanda). It mainly consists of algal flat deposit. The detrital sedimentation of the SD (R2.2) changed towards stromatolitic biostromes. These contain intercalations of nodules (similar conditions to the RSC-SD base), followed by laminites (similar as RSF).

3.1.5 Shituru
The mine of Shituru consists of Mwashya rocks that occur as an anticlinal structure. In the central part of the anticlinal structure an axial breccia can be found. Only in the extreme eastern part of the quarry a bloc of the SDM can be found.

The mineralisation is, however, omnipresent in the lower Mwashya and more specifically in the lower dolomitic level. The typical rocks of the Mwashya at Shituru - with the pyroclastic rocks - have historically been called the "roches or brèches de Kipoi". The mineralisation is interpreted to be part syngenetic and related to a volcanic event. A second phase, perhaps accompanied by an enrichment, very likely resulted from later volcanic hydrothermal activity (Lefebvre 1974). The economic interest of this deposit is formed by the remarkable enrichment, resulting from an intense meteoric alteration.

At Shituru, there are two main orebodies hosted in dolomite of the Kansuki formation, which is located along the southern flank of a northverging syncline with axial plane dipping ~55°SSW. The mined material consists of oxidised ores (10.5wt% Cu), whereas deeper undeveloped primary sulphides with grade ~0.3 to 2.1wt% Cu and 0.05 to 0.1wt% Co have been identified. The Fe-Cu-Co mineralisation includes an early stage of fine-grained disseminated stratiform sulphides (locally framboidal pyrite, chalcopyrite and bornite), with a second stage of bornite, carrolite and linnaeite. The mineralisation is mostly hosted in a partly silicified or recrystallized dolomite and in interbedded dolomitic shale. Host-rock lithologies and sulphide distribution/paragenesis are similar to those of the standard lower and upper orebodies in the Mines subgroup.

Tilwezembe and Mutanda ya Mukonkota, respectively 30 and 45 km ESE of Kolwezi along the same tectonic lineament as Shituru, are two recently explored and mined Cu-Co deposits hosted in rocks of the Kansuki formation. The ores form irregular, supergene bodies hosted in weathered rocks. Brecciated rocks in the vicinity indicate the presence of a tectonic lineament. Grades of the oxide ore are as high as 9wt% Cu and 4wt% Co at Tilwizembe.

The age of the Kansuki volcanic event has been determined at 760+/−5 Ma in the Zambian Copperbelt, at 765+/−5 Ma on mafic flows in Mwinilunga area and at 745+/−7.8 and 752.6 +/−8.6 Ma on metagabbros within upper Roan rocks.

3.1.6 Fungurume
The mineralisation at Fungurume is present in dolomitic formations that consist of an alternance of semi-competent and competent beds (see Oosterbosch 1954). According to this author, the distribution of the mineralisation is related to the deformation of the rocks. The primary sulphides consist of bornite and carrolite and appear to replace the dolomitic host-rock. The bornite is partly replaced by two types of chalcocite, namely
white and blue. Mineralization and silicification are closely related, but different in details. There is no evidence of association of the mineralization with magmatic rocks. The author weighs the evidence for a syn-genetic and epigenetic origin of the mineralisation.

Stratigraphy
In the Fungurume area, the SDM is situated between the RAT and the RGS, i.e. two less competent bancs that are more arenaceous. The SDM exists of:
- Dolomitic limestone (CMN 70m)
- Dolomitic shales (SD 120m)
- Siliceous dolomites, with a massive coarse texture (RSC 20m)
- Laminated siliceous dolomites (RSF 5m)
- Stratified dolomites, fine grain size (DStrat 3.50m)
- Dolomitic chlorite-rich arenaceous shales (RAT gris 0.20m)

The regional deformation has resulted in a succession of regular synclines that are more or less made up of rocks belonging to the Mwashya and Kundelungu. In the anticlinal zones, which are generally faulted and even overthrusted, rocks belonging to the Series des Mines occur as fragments. These are surrounded by brecciated rocks that consist of fragments of RAT or RGS.

The upper mineralisation is present at the base of the SD and has a thickness of several meters. The lower orebody is present in the RSF and DStrat. Oosterbosch (1954) described the occurrence of the mineralisation in the two lower orebodies:

* Upper orebody
  - **F2b**: laminated dolomitic and mica-rich siliceous shales. 1.40 – 3.30m
    These rocks consist of dolomite and quartz, with minor muscovite and chlorite flakes. Typical is the presence of crystalloblasts of dolomite that overlap the borders between successive bands. This dolomite is being replaced by cryptomorphic/phaneromorphic quartz. The mineralisation is present along the margin of mica bands parallel to the bedding, or associated with quartz grains.
  - **F2a**: microgranular siliceous dolomites, with irregular interstratified bands of shale. There is the observation that dolomite is present with an overgrowth of quartz. The sulphides are predominantly associated with this quartz.
  - **F1**: finely grained siliceous dolomites, with mica-rich shales. Microcrystalline quartz is described. 0.50 – 1.75 m
    The sulphides are associated with this quartz. At the base of F1, cherty material replaces dolomite. The mineralisation is present in cracks in the cherts that are filled with quartz or in small veins.

* Lower group
  - **R5** consist of chlorite and dolomite rich shales, in general with
a black colour. The stratification is illustrated by the presence of small chlorite layers parallel to the bedding. Sometimes, the stratigraphy is accentuated by small dolomitic veins of 1 – 2 mm wide, with comb structure. In these veins, the sulphides are overgrown on the dolomite. The intercalation of schist lenses is strongly fractured, with the impregnation of sulphides along these fractures.

- **R4** consists of fine-grained dolomites, interbedded with dolomite or chert nodules. This horizon is characterised by the presence of quartz/chert nodules. In addition, grains of dolomite are immersed in the quartz-rich matrix of the rocks. The core of the nodules consists of dolomite, while the margin consists of dolomite and quartz. The mineralisation is found associated with the margin of the nodules. Some nodules are replaced by black chert. This replacement has advanced from the core to the centre. This horizon shows small interstratified dolomite veins and zones with a high silicification.

- **R3** chloritic dolomitic shale

Similar as R5

- **R2** consists of a finely grained light-coloured dolomite that is in general more or less massive and contains quartz-rich lenses. The dolomites also contain cherty inclusions. Scarce mineralisation

- **R1** contains crystalline dolomites that is more or less nicely stratified. Some bands of cherts can be found. Large dolomite crystals. The mineralisation is present in the dolomites.

The SDM passes into an arenaceous formation, into a shale formation and then in a breccia of RAT rocks. In some places, the breccia is at direct contact with the mineralised SDM. The sulphide mineralisation occurs as total or partly replacement or as a system of small lenticular veins oblique to the stratification. This system is interpreted as a system of shear joints, which is conjugated with the longitudinal shearing responsible for the foliation microstructure.

The RSC forms the border between the two sections with mineralisation, is about 20m thick and made up of algae. It has a heterogeneous primary structure. It was subjected to dolomitisation followed by silicification, with preservation of the microstructure.

The dolomites are interpreted as primary and is only observed in the R1. It is postulated that the chlorite in the other layers hampered the dolomitisation. Silicification is present in three forms: cryptomorphic in the matrix of the shales and the dolomites; massive chert; a phaneromorphic form that is associated with macrocrystalline dolomite and contains sulphide inclusions. There seems to be no relationship with the amount of silica and the tectonic deformation. It must be inferred that the differences in silica content
existed prior to the deformations, either during sedimentation or diagenesis. The chert is interpreted to have formed prior to the mineralisation.

The oxidised zone of the mineralisation has been described down to 100m, followed by a mixed zone until 250m. The orebody has been encountered by boreholes until a depth of 990m. Signs of oxidation have been found until this depth. Chalcopyrite is practically absent. It rarely occurs in association with bornite and the textural relations of the minerals appear to result from unmixing and not from replacement. The bornite occurs as overgrowths of carrolite. It is interpreted that the chalcocite is an overgrowth on both the bornite as the carrolite. The carrolite/bornite is associated with quartz. Also digenite (bleu chalcocite?) and covellite has been described. The grade of the upper sulphide zone, in which chalcocite dominates, can be 6 to 8%. The superficial oxidation makes the grade drop to an average of 5.5 to 6%. In the deep zones where bornite dominates grades vary between 7 and 5.3%. Cobalt grade in general does not exceed 0.4%. It occurs as carrolite, sphaerocobaltite or black oxides.

Oosterbosch (1954) makes a distinction between a possible syngenetic – epigenetic origin for the Fungurume mineralisation.

* syn-genetic based on :
  - the stratified structure
  - the absence of feeding channels
  - the absence of mineralisation in certain formations
  - the uniformity of the mineralisation in 3D
  - the disseminated nature of the mineralisation

* epigenetic based on :
  - the vein-like nature part of mineralisation
  - the replacement of dolomite by sulphides
  - the high temperature forms of bornite
  - the non-existence or slight indication of cataclastic deformation
  - the relationship of the mineralisation to the major tectonic events.

3.2 Eastern part of the Copperbelt

3.2.1 Etoile
In the Cu-Co mineralisation of Etoile (Figure 10), the mineralisation is present in the central part of the Roan Group. The lower part of this sequence is characterised by the presence of volcano-sedimentary rocks, which are associated with a mineralisation. According to Lefebvre & Cailteux (1975), the mineralisation is syn-genetic/diagenetic in the sedimentary rocks (D-Strat) and diagenetic in the pyroclastics rocks (RAT gris).
The total package from CMN to RAT gris is ~120m thick. The thickness of the Series des Mines is not even the half of the same stratigraphic section in the west (Kamoto deposit). There is a breccia at the base of the RAT gris and in the RAT lilas. This brecciated part does not always occur at the same level. Possible volcanic rocks are observed at the top of the RAT lilas and at the base of the RAT gris. The RAT gris consist of a totally different type of rock that is normally observed at this stratigraphic level. According to Lefebvre & Cailteux (1975), it has been interpreted as a sub-marine deposit of volcanic ash. Also at the base of the RSF, iron-rich pyroclastic rocks are observed.

The Roan rocks are present in an anticlinal structure, along which the southern flank has shifted across the northern flank. This resulted in lower Roan rocks thrust on top of Kundelungu rocks. On the cross-sections, it seems that the base of the thrusted rocks belong to the RAT lilas. A part of this RAT lilas is still in stratigraphic sequence.

-CMN consists of an alternance of impure dolomites and dolomitic pelites, with quartz and feldspars. The upper part of the CMN consists of magnesium-rich light coloured dolomites, while the lower part consists of organic material-rich and mica-rich dolomites. The dolomites are finely laminated and consist of dark-coloured bands with quartz and detrital biotite and light-coloured bands with dolomite. The morphology of the CMN is similar to that for the central part of the Copperbelt.

According to Lefebvre & Cailteux (1975), the diagenetic minerals are tourmaline (in the dolomitic layers), microcline (in the sedimentary layers) and quartz.
-SDS: In analogy to other rocks in the western part of Katanga, the contact between CMN and SDS is placed at the top of an organic-rich sedimentary horizon. The SDS consists of an alternance of pure dolomites and feldspar-rich sandy pelites that contain little dolomite. Quartz overgrowths on quartz crystals and microcline overgrowths on K-Spar have been identified as diagenetic alteration. Framboidal pyrite can be found associated with the transformation of ilmenite in leucoxene and leucoxene-rutile. At some horizons in the SDS, regular bancs parallel to the stratigraphy consist of cherty material. The presence of detrital tourmaline, mica and K-feldspar are described.

-BOMZ: The transition between the SDS and the BOMZ is characterised by the appearance of large volume of dolomite associated with quartz, but mostly by the disappearance of the microcline. The level consists of a well recrystallized, badly stratified dolomite. The dolomite is sub-idiomorphic and shows growth zones, along which opaque minerals are associated. The quartz has been recrystallised. There is pyrite mineralisation in the BOMZ. The pyrite is interpreted to be the result of the recrystallisation of frambooidal pyrite. Inclusions of chalcopyrite and galena are found trapped in the pyrite crystals.

-SDB: These are dolomitic rocks, with an average grain size and fine, irregular, stratification, accentuated by the presence of organic material. The dolomites are rich of detrital micas in a chaotic order. No feldspars have been found. The quartz contains numerous inclusions of dolomite. According to Lefebvre & Cailteux (1975), this quartz has been interpreted as diagenetic. Quartz is abundant, but irregularly distributed.

In the entire section, nodules (< 1cm) are found. These have an irregular shape and are often deformed. Due to their presence, they make the stratification go up and down. The nodules are composed of dolomite and quartz, with at the margin chalcopyrite and carrolite.

At Etoile, the SDB is finely mineralised of pyrite that is distributed through the bulk-rock. At some locations in boreholes, the chalcopyrite and carrolite occurs associated with nodules. The chalcopyrite and carrolite overgrows the pyrite. Based on the photographs in the article of Lefebvre & Cailteux (1975), the carrolite formed after the chalcopyrite. Small veinlets that consist of dolomite, and minor quartz, crosscut the dolomitic rocks. They also contain the association chalcopyrite and carrolite. The chalcopyrite seems to overgrow the carrolite crystals. At certain locations, frambooidal chalcopyrite is observed and is interpreted to be due to the replacement of previous frambooidal pyrite. Secondary enrichment locally resulted in the formation of covellite and digenite.

-RSC: Ensemble of massive grey dolomite that is coarsely recrystallized, with often large black crystals of dolomite. The body of the RSC is mostly formed by a stromatolitic bioherm, of which the interstitial space is filled with pelitic material. The rocks have been submitted to a dolomitisation and intense silicification. The neoformation of apatite and rutile is described. Detrital muscovite is present. The
mineralisation is rare and disseminated. It consists of the association of chalcopyrite and carrolite, with digenite. Below the RSC, the rocks have been tectonically disturbed, which disturbed the successions from borehole to borehole. The inferior horizons also show a rapid lateral variation.

-RSF: The transition form RSC to lower horizons passes abruptly with a finely stratified rock. The rocks are dark coloured and look petrographically similar to the RSC due to the dolomitisation and silicification. In comparison to the RSF in the western and central part of the Copperbelt, there is much more detrital muscovite in the organic rich layers. Apatite and tourmaline are recognised as diagenetic minerals. The mineralisation is similar as observed in the RSC. In the lowest part, a chalcocite-rich layer can be observed. This has been interpreted as the result of superficial weathering. As inclusions in this chalcocite, remnants of digenite, carrolite and chalcopyrite are found back. This lower horizon is very different from the rest of the RSF. It is more massive and consists of round quartz/chalcedony fragments that are connected by a quartz/chalcedony cement. It has been described to be the result of the devitrification of volcanic material (Lefebvre & Cailteux 1975). The transition to the DStrat is made by 50cm of banded rocks that consists mainly of hematite, partly transformed to goethite. This hematite enters the rocks as fine veinlets. Small flakes of white mica and rutile crystals are found in these rocks.

-DStrat. The section of D-Strat in the investigated boreholes is abnormally thin compared to the typical sections. There is about two meters of beige-coloured stratified dolomite. The mineralisation largely consists of carrolite that is finely dispersed in the sediments. In joints, the carrolite is more abundant and is associated with chalcopyrite. The chalcopyrite is more limited to certain layers and shows some exsolution laminae of bornite. The rocks consist in general of finely grained dolomite, with little quartz. Chalcedony-rich zones can be found in this dolomite. Quartz-rich crystals overgrow these chalcedony fragments. This larger grained quartz also contains inclusions of dolomite, tourmaline and apatite. In this layer the mineralisation is present as chalcocite with a rim of carrolite. At the east of the deposit, the DStrat can be divided into two parts. The upper part consists of a dark coloured dolomite with large nodules. In the centre of the nodules, there are some indications of chertification. This chertification is followed by the precipitation of quartz and dolomite that is associated with carrolite. There are inclusions of dolomite and chert in the carrolite. The carrolite is crosscut by a later dolomite generations. There is little chalcopyrite associated with the carrolite. Supergene processes resulted in the formation of covellite and digenite and a secondary Co-sulphide. Outside the nodules, the rocks are finely grained, with detrital muscovite and quartz. The lower part of the DStrat is crème coloured and does not contain any nodules. The rocks are finely grained and contains diagenetic quartz with inclusions of dolomite. Feldspar crystals can be found. Microcline has been observed to be associated with pyrite (Lefebvre & Cailteux (1975).
- RAT gris and RAT lilas. In the southern part of Etoile, the RAT is very homogeneous and more than seven meters thick. At the base, the rocks are very laminated and tectonically disturbed. This ensemble is also called the “Formation à minéraux vert (FMV)”. It consists of a green-coloured rock with silica-rich lenses, with chlorite and zeolite. There is no detrital quartz and feldspar in these rocks. The mineralisation consists of chalcocite in the chlorite-rich matrix. In the lenses, chalcocite with cuprite has been identified (Lefebvre & Cailteux (1975). The mineralisation stops abrupt at a brecciated zone. This brecciated zone has a thickness of one meter and is imposed on lila-coloured dolomites (RAT lilas). In the RAT lilas, round fragments can be found that are rich of iron and titanium. These are interpreted to be of volcanic origin. Below the breccia of RAT lilas, rocks of the Ku are found.

Authigenic crystallisation of microcline is described at different stratigraphic horizons. It occurs as overgrows on microcline or as neoformation with numerous inclusions. In the D-Strat, microcline is described to overgrow Fe- and Cu-sulphides (Lefebvre & Cailteux, 1975). The microcline is cut by small fractures formed due to syn-sedimentary faulting, which favours an early diagenetic precipitation of the sulphides.

The precipitation of the quartz is described to occur later in the diagenesis. This is especially the case for the nodules and the lenses, where the quartz is associated with recrystallised dolomite and Cu-sulphides. Other minerals found are tourmaline, muscovite, dolomite, chlorite and rutile. The composition of veinlets and the nodules is the same. They consist of quartz and dolomite with carrolite and chalcopyrite. The dolomite is often coarsely crystalline and hypidiomorphic. The chalcopyrite is present in the entire veins, while it is only present in the pockets in the surrounding rocks.

It should be noted that veins containing quartz, diophane, corundum, iron hydroxides have been described (borehole 283) in the RAT gris. The rocks in the vicinity of these veins show an alteration with the formation of quartz and tourmaline. A similar observation has been described in the SD of Luishya (de Magnée, 1941). Only a low-grade metamorphism has been described (muscovite-chlorite). In the western part of Étoile, metamorphic albite has been described in the RAT gris.

3.2.2 Kipapila

As in Étoile, the RAT gris consists of volcano-sedimentary rocks. The Kipapila deposit is situated in the same anticlinal structures as the deposits of Ruashi and Étoile (Figure 11). The mineralised beds of Kipapila have been either interpreted to have a Mwashya age due to the presence of graphite-rich schistes and granitoid arkoses or to belong to the Series des Mines based on the succession of graphite-rich schistes and the dolomite-rich schistes, with a stromatolitic layer below (M’baya – Kapapa versus RSC).
Different successions of rocks have been identified at Kipapila:

- “Type lithologique 1” consists of layered, organic-rich, very dolomitic material that has a fine to average grain size. There is detrital quartz with diagenetic overgrowths, detrital biotite, dolomite, tourmaline, chlorite and apatite. Pyrite and chalcopyrite occur disseminated along the stratification. The tourmaline is interpreted to have a diagenetic origin.

- “Type lithologique 2” consists of a massive grey-coloured rocks consisting of phengite and dolomite. There is also detrital quartz, muscovite and tourmaline. In addition to phengite, the metamorphic minerals are chlorite, quartz and apatite. There is some pyrite. The contact with the following “Type lithologique” consists of a brecciated rock. This heterogeneous breccia consists of fragments of dolomite, phyllites, quartz and authigenic phengite and phlogopite.

- “Type lithologique 3” consists of a massive green-coloured rock of which the macroscopic appearance looks similar like the RAT gris rocks observed in Ruashi and Etoile. The base of this “Type lithologique” consists of brecciated dolomite, with little calcite, while the upper part consists of silicified dolomite. The central
part consists of clay-rich rocks that are slightly silicified. A mixture of vermiculite/chlorite is described. The quartz in the upper part is interpreted to be formed due to the recrystalisation of the rocks and contains inclusions of dolomite, chlorite, apatite and tourmaline. There is chalcopyrite, which can locally be altered to bornite and digenite at its margins.

- “Type lithologique 4” is nicely stratified and coarser-grained with lenses and cherty pockets. According to Cailteux and Lefebvre (1985), it corresponds to the DStrat. It dominantly consists of quartz and dolomite. The quartz is authigenic and contains dolomitic inclusions. The lenses contain chert and little dolomite. The chalcopyrite has been remobilised by metamorphism. It contains rare exsolutions of bornite and numerous inclusions of quartz, dolomite and phengite. A cobalt sulphide has been observed in the centre of a chalcopyrite crystal.

- “Type lithologique 5” is finely stratified, mostly with chert, and slightly organic-rich (type RSF). Detrital muscovite, biotite, tourmaline have been described. The chalcopyrite and pyrite are finely disseminated and associated with leucoxene and rutile. Sometimes the chalcopyrite is coarser grained and at its margin enriched in bornite, digenite and djurleite.

- “Type lithologique 6” is nicely stratified, more detrital and contains quartz-dolomite nodules. The matrix is organic-rich (Type SD de base). The quartz is partly detrital, while it also forms a large part of the nodules. There are some quartz veins that crosscut the rocks. There is also detrital tourmaline. Metamorphic phengite is present. The nodules contain pyrite at there margin. A fine dissemination of pyrite and chalcopyrite is present in the rocks.

- “Type lithologique 7” consists of massive bancs of macrocrystalline dolomite (type BOMZ). Metamorphic phengite and white mica is present. Pyrite is present with inclusions of dolomite. Nicely banded graphite-layers are found in this lithology. This rock consists of detrital mica and quartz and small quartz-rich veins with little dolomite. Pyrite is finely disseminated.

- “Type lithologique 8” is dominantly arkose-rich, with little dolomite. The detrital fragments consist of quartz, muscovite, sercite, biotite, green tourmaline and microcline fragments. Diagenesis resulted mainly in recrystalisation rims around quartz, microcline and tourmaline. The detrital horizons make this succession correspond to the SDS. Pyrite occurs nicely disseminated throughout the rocks, associated with ilmenite-rutile. In the quartz-dolomite veins, the pyrite is locally accompanied by chalcopyrite.

- “Type lithologique 9” starts with a pure dolomite. The detrital components consist of quartz and microcline. The rocks are characterised by a potassic alteration that consist of a growth rim around the feldspars. Monazite, apatite and phlogopite are described. This section has been attributed to the CMN.
It should be noted that there is a brecciated section between “Type lithologique 2 and 3” and above type 9. Type 1 and 2 have been attributed to upper part of the Dipeta (Cailteux and Lefebvre, 1985).

Due to the absence of detrital material and chlorite, the RAT gris has been interpreted to have a volcanic origin.

### 3.2.3 Hematite in the Lubumbashi area

Important hematite veins can be found in the Mwashya and the Grand conglomerat Subgroups (Buffard and Muhagaze, 1981). These veins can be found in the anticlines of Kifumazi, Ruashi, Lupoto-Kisanga and Kipushi. The veins consist mainly of quartz and hematite, occasionally dolomite. The hematitic veins of the upper Mwashya in Lubumbashi have always been considered to be of hydrothermal origin. However, they are reinterpreted by Buffard and Muhagaze (1981) to be the result of a regional low-temperature metamorphism on chemical and detrital horizons, rich of iron oxides. These horizons are associated with alternating sequences due to seasonal changes or with interfaces localised along the main hiatuses (vein layers). The hematite veins may also be the result of the filling “per decensum” of a fissure network (transversal veins). The observed modifications are silicification, talcification or a renewed remobilisation of in situ of goethite into hematite. In the fissures, the modifications are accompanied by a migration of the oxides from the walls of the veins to the centre of the veins, where they form well crystallized large pockets.

The surrounding Mwashya rock of the hematite rich veins consist of:
- Feldspathic sandstone of ~ 30m
- Dark schistes of ~50-100m
- Green dolomite-rich schistes of ~100m
- The small conglomerat of Mwashya (~1m)

### 4. Database Cu-Co mineralisation in the Katanga fold-and-thrust belt

A systematic overview has been made of all documented Cu-Co mineral occurrences in the UMHK concession of the Lufilian belt (Dewaele et al. 2006, Luyten 2004, Vets 2005). Mineral occurrences with another metal content (e.g. U, Zn, Pb, Ge, Au) have been included in the data (e.g. Cailteux, 1997; Loris, 1996; Meneghel, 1981). The study area was based on the available data and the existence of detailed geological maps (Demesmaecker et al., 1962; François, 1973; 1980; Lagmouch et al., 2004; Oosterbosch, 1962). Data on 230 mineral occurrences were collected from the archives of the Royal Museum of Central Africa (R.M.C.A.) and were compiled with detailed information on each of the localities described. From each mineral occurrence or ore deposit, the coordinates of the mineralisation, the host-rock, the type of deposit (i.e. stratiform, vein-type, enrichment), type of metals, concentration (Mt), estimated reserves, structural setting and proposed genesis are summarised.
Table 1. Overview of the different mineral occurrences in the Congolese Copperbelt. The mineral occurrences are grouped according to deposit type, hostrock and the economic significance of the deposit. R = deposit in Roan Group, R-x = Roan Sub-Group, Ng = Nguba, Ku = Kundelungu, ? = unknown age of hostrock. Mineral occurrences with a different metal content as Cu and Co-like U, Ni, Pb, Cd, Ge, Ba, Au, Fe, Mn, Mo, Re, V, Pd- are grouped under “other” (François, 1973).

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The classification of the deposit in stratiform, enrichment and vein type is adopted from François (1973). However, it must be noted that stratiform mineralisation could have been submitted to supergene enrichment, while supergene mineralisation could have a stratiform occurrence. In this study, supergene mineral occurrences are solely considered as the result of late enrichment and mobilisation and, therefore, have different characteristics. They consist predominantly of Cu-oxides (e.g. cuprite), Cu-carbonates (e.g. malachite) and Cu-silicates (e.g. chrysocolla), while the stratiform and vein-type are a sulphide mineralisation. Mineralisation without any information about their occurrence and characteristics in literature are classified as “unknown”. A mineralisation is classified as economically significant if a mine or a potential exploitation is mentioned in literature (François, 1973).

Table 1 shows an overview of the results obtained from the archives. The mineral occurrences are grouped according to the deposit type, the host-rock and the economic significance of the deposit. Stratiform mineralisation forms the largest group (169 of 230). Stratiform mineralisation is hosted predominantly in rocks belonging to R-2 (142 of 169 deposits, Table 1). Only 27 of 169 stratiform occurrences can be found in rocks belonging to the R-3, R-4, Nguba or Kundelungu. The second largest group results from supergene enrichment (38 occurrences). The “enrichment” deposits are present in sediments that range in stratigraphic occurrence from R-2 to Kundelungu. Deposits that formed by enrichment are also mainly present in Roan rocks (25 of 38 deposits). Only 5 of 36 occurrences of the enrichment type are located in R-2. The third group with 8 occurrences consists of vein-type mineralisation and includes deposits such as Kipushi, Shinkolobwe and Swambo. Finally, 15 mineral occurrences could not be classified and are termed unknown.

Based on this systematic overview of mineral occurrences in the Congolese part of the Lufilian fold-and-thrust belt in Katanga, it can be concluded that the majority of mineralisation is stratiform and occurs in R-2 rocks, while mineralisation in rocks younger than R-2, is mainly due to the remobilisation of existing primary mineralisation by supergene processes (Dewaele et al. 2006).

5. Satellite imagery and mineralisation

Satellite images have been interpreted to identify geological structures and lineaments. Lineaments have been defined as “rectilinear or gently curved alignments of topographic features on a regional scale, generally judged to reflect crustal structures” (Hobbs, 1911). Structural and morphological elements are combined into lineaments of composite nature. During this study, LANDSAT images have been used with a spatial resolution of 28.5 * 28.5 m² /pixel. The data have been integrated in a Geographic Information System (G.I.S., MapInfo Professional Version 5.5). Layers with the location and characteristics of the mineral deposits, the geological map and the interpreted structural features have been combined to identify relationships between mineralisation, stratigraphy and geological structures. On the satellite images, ridges could easily be identified. These ridges are formed by rocks with a high resistance to erosion. By combining the layer of the satellite images with the layer of the geological map, it is possible to relate these ridges with a certain stratigraphical interval and structural features. Ridges can be
identified along the margins of thrust sheets and along faults in these thrust sheets and consist mainly of R-2 sediments. The ridges in the central part of the thrust sheets consist dominantly of Nguba (Ng)- or Kundelungu (Ku)-Group rocks (Dewaele et al. 2006, Vets 2004, Luyten 2005).

In addition, by combining the satellite images with the layer with the description and location of the mineral occurrences, a clear relationship can be observed between the ridges and Cu-Co mineralisation. With the exception of the Kolwezi and Mashitu regions, the stratiform mineralisation can be related with ridges along the margins of thrust sheets and along faults in these thrust sheets, since these consist mainly of rocks belonging to the R-2 Subgroup. Although the Kolwezi klippe and the Mashitu area are intensely mineralised, no geological structures can be identified. However, this could be due to the intense mining activity, with the omnipresence of mining dumps and/or the presence of strongly deformed Roan sediments. Only minor stratiform mineralisation is found associated with the ridges in the central part of the thrust sheets. However, the lack is not surprising since Roan rocks are missing. Four stratiform mineralisation are identified inside the thrust sheets (Kabulo North, Kafumvua South, Musoko North and Luwowishi; François, 1973; 1974), which occur in Kundelungu rocks. They are described as stratiform deposits (François, 1974) and are associated with lineaments. Malachite has been described for these locations.

Enrichment mineralisation can be found along ridges in the central part of thrust sheets, along their margins and along faults in the thrust sheets. The majority of enrichment mineralisation within the R-3-4, Nguba and Kundelungu and often also in the R-2 are located along lineaments. For example, in the Mukinga area, the stratiform R-2 mineralisation is located along ridges of Roan rocks, while the Mukinga-south deposit, present in the Nguba-Group and described as a mineralisation that formed by enrichment, occurs along a small lineament with a NE-SW orientation. The lineaments could have formed pathways for the remobilising fluids that formed the mineral occurrences in the
Roan 3-4, Nguba and Kundelungu strata. Also, 7 of the 8 vein-type deposits are related to lineaments. However, for the Swambo mineralisation in R-2 rocks no lineament has been identified.

We investigated the potential relationship between the occurrence of mineralisation and lineaments (e.g. Shituri, Figure 12). This is often difficult due to the cover by products of mining activities. The majority of the described economic important ore deposits are, however, related to lineaments (Table 2). A similar observation can be even made for the major stratiform ore deposits belonging to the R-2 group. Economic stratiform Cu-Co mineralisation are situated at the intersection of R-2 strata and a crosscutting lineament. The latter could be interpreted as fractures zones/faults along which mineralising descended or ascended.

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Table 2. Relationship between mineralisation and potential lineaments.
6. Petrography of stratiform mineral deposits

6.1 Western part of the Copperbelt

Numerous deposits occur in the western part of the Copperbelt, but the deposits situated on the Kolwezi Klippe are the most important. For this study, the Kamoto and Musonoi deposits have been selected for a more detailed investigation. The paragenetic evolution of the Cu-Co stratiform mineralisation at the Kamoto Principal and Musonoi mines has been studied in detail. Musonoi is located at a few kilometres west from the well-known Kamoto mineralisation (Bartholomé et al., 1972; Bartholomé, 1974). Both deposits have a km-scale extension. A continuous sampling in the R-1.4 to R-2.3 was possible thanks to the in-house collection and thanks to the collection of P. Bartholomé at the University of Liege (Belgium).

During this study, only rocks belonging to R-2.1 and the lower part of R-2.2 have been studied petrographically (Figure 13). The Kamoto and Musonoi mineral occurrences consist of a lower and upper orebody, separated by a low-grade stromatolitic dolomite (R.S.C.: “Roches Siliceuses Cellulaires”). The lithology of the “lower” orebody host-rock consists of a chloritic-dolomitic siltite, stratified and silicified dolomites, while the host-rock of the “upper” orebody consists mainly of dolomitic shales, dolomites and fine-grained sandstones. Bedding shows millimetre to centimetre thicknesses, which reflect grain-size and mineralogical variations (François, 1973).

![Figure 13. Detailed stratigraphic cross-section of the mineralised parts of the Series des Mines.](image-url)
Polished sections have been studied by transmitted and incident light microscopy and cold cathodoluminescence petrography. A Technosyn Cold Cathode Luminescence Model 8200 MK II was operated at 10-16 kV, 5.5 Pa, 5 mm beam width and a current below 1000µA. The host-rock consists mainly of detrital dolomite, quartz and muscovite with some minor amounts of chlorite and tourmaline. The detrital dolomite and quartz are strongly compacted and the muscovite grains are preferentially oriented along the bedding plane. Framboidal pyrite is only identified in the upper part of the upper orebody (R-2.2.1.2). Nodules, consisting of anhydrite and gypsum formed in the evaporitic environment and an early dolomitisation, with non-luminescent dolomite, occurred.

An important part in the stratiform mineralisation is played by anhydrite pseudomorphs (Muchez et al. 2008). Pseudomorphs after evaporates occur in dolomites and dolomitic siltstones of the Mines Subgroup in the Katangan Copperbelt (Cailteux 1977, 1994; Lefebvre 1978). The host rock comprises fine-grained xenotopic dolomite (µm size crystals) or coarser grained hypidiortopic dolomite (20 - 300 µm). Locally, the fine-grained dolomite recrystallised to a bright yellow-orange luminescent dolomite, which also occurs as a cement postdating the opaque minerals in large cavities (up to several
hundreds of µm) that form a porous network. Dispersed, framboidal pyrites occur in the fine-grained rocks, reflecting typical products of the activity of sulphate-reducing bacteria in anaerobic sediments (Machel 2001). This reaction takes place at temperatures below 80°C (Machel and Foght 2000). The framboidal pyrites are replaced and overgrown by chalcopyrite (see also Bartholomé et al. 1972; Brown and Chartrand 1983). Disseminated carrolite, bornite and chalcocite occur associated with chalcopyrite. The crystal size of these sulphides mainly varies between 20-150 µm, similar to the size of the framboidal pyrites. In addition, lenses of sulphides may reach one mm in thickness.

The mineralised rocks are characterised by the precipitation of authigenic quartz, associated with the mineralisation (Dewaele et al., 2006). The authigenic quartz may occur disseminated, in clusters or massive and may replace completely the original laminated dolomitic siltstone. Detrital quartz and dolomite are absent in the silicified rock. However, randomly oriented phyllosilicates are still preserved. The phyllosilicates in this authigenic quartz are randomly oriented and may show a rectangular structure. Pyrite occurs both in the laminated host rock and the silicified rock, but the grain size is much larger in the latter than in the original rock. This indicates remobilisation of sulphides during the formation of the authigenic quartz. Chalcopyrite that occurs disseminated in the laminated rock can also occur concentrated along the reaction front of authigenic quartz. Dissolution of detrital quartz may also be restricted to more coarse-grained layers, which is reflected in the presence of coarse-grained phyllosilicates and the precipitation of authigenic quartz in clusters.

Pseudomorphs after anhydrite are abundant in the Mines Subgroup (e.g. Cailteux 1977, 1994; Lefebvre 1978). They occur in oval to round and cauliflower nodules, in lenses that maybe wedge shaped and in lath-shaped crystals. In the pseudomorphs, dolomite precipitated first and was replaced by sulphides and authigenic quartz. The fluid inclusions in the authigenic quartz belong tot the H2O-NaCl system, have a salinity between 8.4 and 18.4 eq. wt% NaCl and homogenisation temperatures between 80 and 192°C (Dewaele et al. 2006).

Oval to rounded nodules have a size between a few hundred µm and 5 mm. The nodules consist of a rim of authigenic quartz (AQ) and sulphides, such as chalcopyrite, carrolite and chalcocite. The size of these crystals varies between 20 and 200 µm. The core of the nodules is composed of a brown luminescent hypidiotopic dolomite (up to 1.2 mm). AQ and sulphides replace the dolomite crystals and therefore postdate dolomite formation. Although the sulphides and the authigenic quartz dominantly occur at the rim of the nodule and the dolomite in the core, the nodule may also be completely replaced by authigenic quartz and sulphides with a few relics of dolomite. Cauliflower nodules with a size up to 2 cm consist of a coarse-grained hypidiotopic dolomite (80 – 300µm) with a zoned brown luminescence, AQ and sulphides. The latter two replace the dolomite.

Pseudomorphs after anhydrite are also present as lenses (size ranging between 250 µm to 1 mm width and 1.5 mm - 2.5 cm length). Wedge-shaped lenses are several cm long and 15 mm thick. The lenses are composed of authigenic quartz, bornite, carrolite, chalcopyrite, chalcocite and some relics of brown luminescent dolomite, which is replaced by AQ and sulphides. The AQ often contains inclusions of sulphides, but
sulphides may also overgrow authigenic quartz. Lath-shaped pseudomorphs after anhydrite are ~100 by 400 µm large and are only composed of AQ and dolomite. Carrolite typically occurs in the pseudomorphs after anhydrite and is therefore more abundant in rocks with these pseudomorphs than without, suggesting a genetic relationship.

### 6.2 Eastern part of the Copperbelt

Numerous deposits occur in the western side of the Copperbelt. The Cu deposits extend to the south to the Kimpe area. In this report, only information is given about deposits situated in the Copperbelt until the area of Lubumbashi. The deposit of Luiswishi has been considered as representative for this area. Only a limited number of samples of the primary mineralisation of this deposit are available in the collections of the RMCA. This could implement that the established paragenesis is not complete.

The host-rock of the mineralisation consists mainly of recrystallized dolomite, quartz and muscovite. In rocks of the SD, the presence of nodules can be observed. These nodules consist mainly of quartz, no dolomite has been observed. Only little chalcopyrite mineralisation is associated with these nodules. Mineralisation occurs also disseminated in the rock. Chalcopyrite occurs associated with quartz in a recrystallized dolomitic matrix.

The rocks are strongly fractured. These small fractures are filled with small dolomite crystals. This phase was followed by a main vein generation. Large dolomite crystals formed, followed by the precipitation of quartz. The main Cu-Co mineralisation occurs associated with these large dolomite crystals. The Cu-Co mineralisation dominantly consists of chalcopyrite and carrolite. It seems that the carrolite formed earlier than the chalcopyrite due to mutual overgrowths and the occurrence of chalcopyrite in fractures in the carrolite crystals.

### 5.3 Central part of the Copperbelt

The mines of Kambove, Kamoya and Luishya have been studied as representative for the stratiform mineralisation in the central part of the Congolese Copperbelt.

A continuous set of samples is available from the Kambove west deposit. Samples have been studied from the RAT gris up to the CMN horizon. A distinction can be made between samples that have an undisturbed appearance and samples that exhibit numerous veinlets.

For the undisturbed samples, the observations are quite similar as for Kamoto. The bulk of the rock consists of fine crystalline dolomite, with quartz and muscovite. Mineralisation occurs associated with pseudomorphs of evaporate nodules and disseminated in the host-rock. This first mineralisation stage is associated with authigenic quartz and consists of chalcopyrite, carrolite, bornite, digenite and chalcocite. The first dolomite generation that forms the bulk of the rocks at Kambove is locally recrystallized in a dolomite generation with a larger crystal size. In addition, small veinlets can be found that consist of relatively coarser crystalline dolomite. In these veinlets, carrolite has been observed.
In the tectonically disturbed rocks of Kambove, it can be observed that there was first a penetrative fracturing of the rocks, filled with chalcopyrite and bornite. These small veins also consist of quartz and dolomite. This phase is followed by an important development of large veins with large dolomite, quartz crystals and muscovite. These veins contain also large chalcopyrite and bornite crystals.

Samples have been obtained from different mineralized stratigraphic horizons at the Kamoya mine. Samples have been investigated from the DStrat up to the SDb. The rocks dominantly consist of fine-grained dolomitic rocks with considerable quartz. Typical is the occurrence of nodules that consist of dolomite and authigenic quartz. These nodules are often mineralized and contain chalcopyrite and carrolite. These two sulphides occur also disseminated in the host-rock where they are also associated with authigenic quartz. Locally, the recrystallized dolomites are crosscut by small quartz-minor dolomite veinlets. These can contain bornite and chalcopyrite.

Samples from Luishya come mainly from the “Carrière des Pelles” and the “Carrière B and C”. The host-rock of the mineralisation of Luishya consists mainly of recrystallized dolomite that only contains minor quartz. In one sample, remnants of magnesite crystals have been observed that are completely recrystallized into dolomite. As in Kambove, the fine-grained dolomite is recrystallized in larger dolomite crystals. Chalcopyrite and carrolite are associated with both dolomite generations. The zones of recrystallized dolomite can form locally very rich zones. Small nodules have been observed. They consist of coarse crystalline dolomite and are associated with carrolite and chalcopyrite. No distinction can be made between the carrolite and chalcopyrite that occur associated with the nodules or with the bulk dolomite.

The rock is crosscut by numerous veins. In an initial stage there was a massif recrystallisation/impregnation of the host-rock with quartz and dolomite. This phase is associated with the precipitation of disseminated bornite. It is followed by a main vein generation that contains of large crystals of quartz and dolomite. Large crystals of chalcopyrite and bornite are also present in these veins. Although the mass of chalcopyrite formed prior to bornite, exsolution of chalcopyrite from bornite occurs at the margin of the bornite crystals. Bornite also occurs as a late phase that crosscuts the carrolite and chalcopyrite that occurs in the recrystallized dolomite.
7. Geochemistry

7.1 Sulphur isotope composition

7.1.1 Literature
To provide further constraints on the conditions of the sulphide formation in the stratiform mineralisation, stable isotope studies were carried out on the sulphides. Dechow and Jensen (1965) made a general overview of the sulphur isotope composition of sulphides from different mineral deposits in the Lufilian fold-and-thrust belt in Zambia and the DRC, but also from sulphides occurring in surrounding sedimentary and magmatic/volcanic rocks. They concluded that the sulphur in the stratiform deposits originates from “biogenic sulphate”, i.e. seawater sulphate.

Sweeney et al (1984) and Sweeney & Binda (1989) have done analysis of sulphates/sulphides that occur in different horizons in the Konkola basin (Table 3). The values obtained for anhydrite didn’t show any significant change (+17.8 ‰) at Chambishi, which is close to the value reported for late Proterozoic seawater (+18.9 ‰, Veizer et al. 1980). Therefore, they concluded that the δ^{34}S values they measured for the sulphides in the ore shale are stratigraphically correlated (see also Dechow & Jensen 1965) and mirror the transgressive/regressive events identified based on sedimentary features. Transgressive events are associated with ^{34}S depletion of the sulphides, while regressive events reflect ^{34}S enrichment (Figure 15).
<table>
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Table 3. Overview of the relevant $\delta^{34}$S information available in literature.
McGowan et al. (2003) made a detailed study of the different sulphide generations in the Nchanga mineralisation. He investigated pyrite in the black shale unit, Cu- and Co-sulphides from the shales and arenites from the upper and lower orebody. They observed a clear difference between the mean values of the lower and upper orebody. However, the observed range spans the range of the diagenetic pyrite that has been replaced. The authors interpreted that the diagenetic pyrite was formed by bacteriogenic reduction, while the sulphides from the orebodies were formed by thermochemical reduction, with continuing increase of $\delta^{34}$S in a closing environment. The sulphur source would have been remobilized sulphate from evaporates.

Hoy & Ohmoto (1989) make a calculation about the origin of the sulphur at Kamoto and Musoshi. The data suggest the addition of significant quantities of secondary sulphides having $\delta^{34}$S of ~9%, next to the sulphide due to the bacteriogenic reduction. It was postulated that the additional sulphide was generated at the depositional site by abiogenic sulphate reduction during the introduction of the Cu-bearing fluids.

Lerouge et al. (2005) made a detailed study of carrolite and chalcopyrite that occurs in different horizons of the Series des Mines at Luiswishi. As previous researchers, they observed that there is a variation that is stratigraphically controlled, with variation due to the isotopically open and closed environment during deposition. Marine sulphate is proposed as sulphate source, with only a slight external contribution. The sulphur isotopes between the stratiform and the stockwork sulphides are the same, so a local-scale reworking of the early sulphides is proposed. This reworking was not strong enough to re-equilibrate the $\delta^{34}$S/$\delta^{32}$S under diagenetic/metamorphic conditions.

### 7.1.2 Recent work
Additional analysis have been carried out by Muchez et al. (2008) and El Desouky et al. (in press) of the stratiform mineralisation at Kamoto/Musonoi and Luiswishi. These analyses have been carried out to characterize the processes of mineralisation (bacteriogenic towards thermogenic sulphate reduction) in more detail.

The sulphur isotopic composition of chalcocite, bornite and carrolite from Type 1 nodules and lenses varies between -10.3 and +3.1‰ V-CDT (Muchez et al. 2008). This is range is comparable to the $\delta^{34}$S values reported for disseminated and bedding-parallel chalcopyrite and carrolite at Luiswishi, i.e. between -14.2 and +5.8‰ V-CDT (Lerouge et al. 2005) and for diagenetic pyrite in black shales in the Zambian Copperbelt (McGowan et al. 2003, 2006).

The $\delta^{34}$S values of the Lower Roan anhydrite nodules in the Copperbelt are between +12 and +22‰ V-CDT (Dechow and Jensen 1965). An average $\delta^{34}$S value of +17.5‰ V-CDT is suggested for Late Neoproterozoic seawater (Claypool et al. 1980). Taking this into account, calculated $\Delta$SO$_4$-H$_2$S values for the copper-cobalt sulphides in the evaporite pseudomorphs fall between +14.4 and +27.8 ‰, consistent with fractionation values typical for bacterial sulphate reduction (Ohmoto 1986; Machel et al. 1995). Therefore, the replacement of anhydrite and the consequent precipitation of sulphides is related to BSR occurring in the anhydrite nodules (Muchez et al., 2008).
Additional isotope analysis have been carried out on chalcopyrite, bornite and carrolite from recrystallized lenses, nodules, veins and breccia cement at Luiswishi (El Desoucky et al. in press). Similar δ³⁴S values as for the sulphides in the undisturbed Type 2 nodules and lenses were obtained, i.e. -13.1 to -0.9 V-CDT. Possible interpretations include an in-situ remobilization of the first phase Cu-Co sulfides during the second fluid migration phase (c.f. Lerouge et al., 2005) or mixing between sulfur from thermochemical sulfate reduction (TSR) of evaporite relicts and sulfur from the first phase Cu-Co sulfides (El Desoucky et al. in press).

7.2 Stable carbon and oxygen isotopes

7.2.1 Literature
To provide further constraints on the conditions of the sulphide formation, stable carbon and oxygen isotope studies were carried out on carbonates that formed contemporaneously with the sulphides. Sweeney et al. (1984) and Sweeney & Binda (1989) have done analysis of the different horizons in the Konkola basin. The Footwall samples have values between -4 to -10 ‰ V-PDB for δ¹⁸O and -6 to -9 ‰ V-PDB for δ¹³C. They concluded that the δ¹³C values of Ore shale carbonates (-8 to -22 ‰ V-PDB) indicate that organic matter was oxidised during carbonate formation. It was suggested that δ¹⁸O, and to a lesser extent δ¹³C, values of the carbonate from the Ore shale and the Footwall rocks reflect the influence of both fresh and marine waters.

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<th>δ¹³C (‰ V-PDB)</th>
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<td></td>
<td>Ore shale</td>
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<td>-8.8 to -20.5</td>
</tr>
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<td>upper orebody</td>
<td>11.7 to 16.9</td>
<td>-8.3 to 2.9</td>
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</table>

Table 4. Historical overview of previous stable oxygen and carbon isotope investigation.

In the extensive study of Selley et al. (2005), 369 carbonate samples (whole rock, veins and evaporite nodules/pseudomorphs) were analysed, confirming a trend recognised by earlier workers (Annels 1989; Sweeney and Binda 1989) from values typical of Neoproterozoic marine carbonates (δ¹³C = -4.0 to +4.0‰ V-PDB and δ¹⁸O = -8 to -4 ‰ V-PDB; Veizer and Hoefs 1976; Lindsay et al. 2005) in the unaltered sedimentary carbonates to isotopically light values (δ¹³C = -4.0 to -26‰ V-PDB and δ¹⁸O = -8 to -25‰ V-PDB). Carbon and oxygen isotope data are consistent with the introduction of an organic carbon component at higher temperatures (Selley et al. 2005, McGowan et al. 2006). From this and existing data, it is clear that organic carbon played an important role as a source in the precipitation of the dolomites, but there is no criterion to distinguish early from late oxidation reactions. The wide range of measured oxygen isotope values
could be due to a variation in the precipitation temperature of the dolomite, variation in the isotopic composition of the dolomitising fluid, variable interaction of the ambient fluid with the host rock, or a combination of these processes.

McGowan et al. (2006) made a detailed study of the isotopic composition of the dolomites in the Nchanga area. The $\delta^{13}C$ of dolomites from units above the Upper orebody give $\delta^{13}C$ values of $+1.4$ to $+2.5\%e$, consistent with marine carbon. However, dolomite from the shear zones and the alteration assemblages within the Upper orebody show more negative $\delta^{13}C$ values: $-2.9$ to $-4.0\%e$ and $-5.6$ to $-8.3\%e$ respectively. Shear zone and upper orebody dolomites give a $\delta^{18}O$ of $+11.7$ to $+16.9\%e$, which is lower than the Lower Roan dolomites, which show $\delta^{18}O$ of $+22.4$ to $+23\%e$. This is interpreted to be due to thermochemical reduction of a sulphate- (and metal) rich hydrothermal fluid at the site of mineralisation. It is postulated that no contribution of bacteriogenic sulphide produced during early diagenesis contributed to the ores.

7.2.2 Recent work
The carbon and oxygen isotopic composition of the fine-grained dolomitic host rock and the dolomite in the pseudomorphs has been measured at Kamoto (Muchez et al. 2008). The $\delta^{18}O$ and $\delta^{13}C$ values of the host rock dolomite vary between $-7.8$ and $-10.9\%e$ V-PDB and between $-7.0$ and $+1.3\%e$ V-PDB respectively. The dolomites in the nodules have a $\delta^{18}O$ value between $-10.1$ and $-11.5\%e$ V-PDB %e and a $\delta^{13}C$ between $-2.9$ and $-9.9\%e$ V-PDB (Figure 16).

![Figure 16](image_url)

Figure 16. Carbon and oxygen isotope composition of fine-grained dolomite of the Mines Subgroup and of hypidiotopic dolomite pseudomorphs after anhydrite. Also shown is the isotopic composition of Neoproterozoic marine dolomites (after Muchez et al. 2008).
The isotopic composition of Late Neoproterozoic marine dolomites ranges between -8 and -4 ‰ for oxygen and between -4.0 and +4.0 ‰ for carbon (Veizer and Hoefs 1976; Lindsay et al. 2005). The isotopic composition of the host rock dolomite partly falls in this range. The composition of the dolomite pseudomorphs falls outside this range and lies at the lower end of the range of the host rock dolomites. The lowest $\delta^{18}O$ value of -11.50‰ V-PDB can be explained by dolomite precipitation from seawater at a maximum temperature between 55 and 70°C (cfr O’Neil et al. 1969) and thus, within the temperature range of bacterial sulphate reduction. The low $\delta^{13}C$ values are the result of preferential incorporation of $^{12}C$, generated during the oxidation of organic matter (Irwin et al. 1977).

7.3 Microthermometry

7.3.1 Literature

Previous work on fluid inclusions in the Copperbelt in Zambia and the DRC includes Pirmolin (1970), Sweeney (1987), Richards et al. (1988), Speiser et al. (1995), Kamona (1993), Greyling et al. (2005) and McGowan et al. (2006) (Table 5).

<table>
<thead>
<tr>
<th>Country</th>
<th>Stratigraphy</th>
<th>Mineral</th>
<th>Origin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pirmolin 1970</td>
<td>Kamoto, DRC</td>
<td>Series des Mines</td>
<td>Cherty dolomite</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Epigenetic stratiform</td>
</tr>
<tr>
<td>Sweeney 1987</td>
<td>Konkola, Chambishi, Zambia</td>
<td>Lower Roan</td>
<td>Quartz vein</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Late diagenetic stratiform</td>
</tr>
<tr>
<td>Annels 1989</td>
<td>Chambishi, Zambia</td>
<td>Lower Roan</td>
<td></td>
</tr>
<tr>
<td>Richards et al 1988</td>
<td>Musoshi, DRC</td>
<td>Lower Roan</td>
<td>Quartz vein</td>
</tr>
<tr>
<td>Kamona 1993</td>
<td>Kabwe, Zambia</td>
<td>Lower Kundelungu</td>
<td>Dolomite</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Epigenetic Pb-Zn</td>
</tr>
<tr>
<td>Speiser et al 1995</td>
<td>Kansanshi, Solwezi, Zambia</td>
<td>?</td>
<td>Quartz vein</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Epigenetic iron-oxide-Cu-Au</td>
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<tr>
<td>Kampunzu et al 1998</td>
<td>Kipushi DRC</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Epigenetic</td>
</tr>
<tr>
<td>Greyling et al 2005</td>
<td>Chambishi</td>
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<td>Quartz vein</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Late diagenetic to early metamorphic</td>
</tr>
<tr>
<td>McGowan et al 2006</td>
<td>Nchanga</td>
<td></td>
<td>Quartz veins</td>
</tr>
</tbody>
</table>

Table 5. Overview of previous microthermometric investigation on deposits in the Lufilian fold-and-thrust belt of the DRC and Zambia.

Pirmolin (1970) described fluid inclusions from an unmineralised cherty dolomite layer of the Series des Mines. The dolomite contains inclusions with solid phases. Homogenisation of the L+V phase occurs between 225 and 240°C ~200°C. Calculated salinities based on the dissolution temperature of the solid phase is ~40wt%NaCl.
Crosscutting quartz veins at Konkola and Chambishi were studied by Sweeney (1987). These veins have undergone at least one post-formational tectono-thermal event. SEM analysis indicated the presence of Na, Al, Si, S, U, K, Ca, Fe, Cu and Co and NaCl, KCl daughter minerals. It was concluded that the veins formed by lateral migration of fluids during late diagenetic dewatering.

Annels (1989) cited the fluid inclusion work of Cunningham (1986) on samples from Chambishi and Chambishi SE. These samples were sub-concordant vein quartz and quartz nodules that host sulphides. They revealed homogenisation temperatures of 130-165°C and a salinity of 9-16 eq wt% NaCl for veins associated with early pyrite and carrolite, and 125-145°C and 16-22 wt% NaCl for later copper sulphides (chalcopyrite and bornite).

Richards et al. (1988) indicated the presence of a sylvite-halite saturated fluid in quartz-hematite-rutile veins hosted in the footwall sediments of the Musoshi copper deposit, where late hydrothermal veining caused extensive footwall- and ore shale alteration. Fluid inclusions were found to consist of NaCl (39 wt%), KCl (15 wt%) and minor amounts of CO$_2$, and with a $T_{h_{tot}}$ of ~395°C. This hydrothermal event post-dated stratiform copper mineralisation and is linked to compressional deformation and metamorphism during the Lufilian orogeny.

Epigenetic iron oxide Cu-Au type mineralisation at the Kansanshi copper mine (Solwezi area, NW Zambia) was studied by Speiser et al. (1995) and Speiser (1994). Hydrothermal veining at Kansanshi is connected with alteration of the host-rock and mineralised veins have been dated at 503-511 Ma.

Kamona (1993) studied fluids from the epigenetic stratabound carbonate hosted Pb-Zn deposit of Kabwe. Mineralising fluids have a salinity between 11 and 31 wt. NaCl and $T_{h_{tot}}$ between 257 and 305°C.

Kampunzu et al. (1998) reported high homogenisation temperatures (300°C) from fluid inclusions representative for the Kipushi mineralisation.

Greyling et al. (2005) indicated the presence of an early saline H$_2$O-NaCl+MgCl$_2$+CaCl$_2$+CO$_2$+CH$_4$+N$_2$ fluid (11.9 – 23.1 wt% NaCl) with $T_{h_{tot}}$ between 86-129°C. Later fluids occur in fluid inclusion planes and are interpreted to be linked to early metamorphism of the basin. These fluids are more varied in composition, and are aqueous, aqueo-carbonic, and pure methane inclusions. Aqueous and aqueo-carbonic inclusions are H$_2$O-NaCl+MgCl$_2$+CaCl$_2$+CO$_2$, N$_2$, H$_2$S fluids, with two salinity end-members, i.e. a high saline (18-23 wt% NaCl)-lower temperature (~ 130-160°C), fluid, and a lower salinity (~6 wt% NaCl eq)-higher temperature (~140°-210°C) fluid. Later fluids are thought to be late-diagenetic fluids on the one hand, and early metamorphic fluid on the other.

McGowan et al. (2006) has measured inclusions from quartz veins with carbonate or microcline from the Mines sequence at Nchanga. Veins have been investigated in
extensional structures, in the upper and lower orebody and in thrust-related vein quartz. Three-phase (L+V+S) inclusions have been identified in the different quartz veins. Salinity ranges from ~ 31 to 38 wt% NaCl and homogenisation temperatures vary between 140 and 180°C.

7.3.2 Microthermometric investigation

Recently, microthermometric investigations have been performed by the authors of this report on numerous deposits in the Lufilian fold-and-thrust belt (Kamoto, Musonoi, Dewaele et al. 2006; Luiswishi, El Desoucky et al. 2007; Kipushi, Heijlen et al. 2007, Heijlen et al. submitted) and the Kundelungu foreland (Dikulushi, Dewaele et al. 2006, Haest et al. submitted; Lufukwe, El Desouky et al. in press). In this report, most attention is spent on the results of the stratiform mineralisation of Kamoto/Musonoi, since the fluids explain the most important part of the history of the early diagenetic mineralisation, i.e. early diagenetic precipitation of sulphides that is observed across the entire Copperbelt, from Kamoto to Luiswishi.

Kamoto/Musonoi

Samples were selected to study fluid inclusions in the authigenic quartz, which is interpreted to be contemporaneous with the main stage of the stratiform mineralisation (see also Oosterbosch, 1962; François, 1974). One doubly polished section was prepared from the lower orebody (R-2.1.2; RSF : “Roches Siliceuses Feuilletées”) of the Musonoi mineralisation, and one from both the upper orebody (R-2.2.1; BOMZ : “Black Ore Mineral Zone”) and the low-grade intermediate zone (R-2.1.3; RSC : “Roches Siliceuses Cellulaires”) of the stratiform Kamoto mineralisation. Only quartz has been studied microthermometrically. The microthermometric results obtained by Pirmolin (1970) on dolomite from Kamoto have been interpreted by Okitaudji (1989) to be stretched and, therefore, the homogenisation temperatures cannot be used to deduce ambient precipitation temperatures. Indeed, inclusions in carbonates are easily stretched during subsequent burial or tectonic deformation (cf. Goldstein, 1986; Barker and Goldstein, 1990).

The fluid inclusions measured typically occur as individual inclusions, clustered or concentrated in zones within the authigenic quartz crystals. Some of them occur in growth zones and a primary origin can be proposed. In addition, the inclusions present in the authigenic quartz do not occur in later carbonate generations. At room temperature, all fluid inclusions are two-phase (L + V), while inclusions in younger dolomite generations have multiple daughter crystals (see also Pirmolin, 1970; Ngongo, 1975). Therefore, all inclusions in the quartz are interpreted to predate the precipitation of subsequent carbonate cement generations (cf. Wilkinson, 2001).

Two-phase inclusions have been identified in the authigenic quartz associated with mineralisation in the RSC, the RSF and the BOMZ. The association of authigenic quartz with mineralisation occurs in evaporate nodules and disseminated in the host-rock. Due to the small size of the inclusions (≤ 5 µm), the eutectic melting was only visible in a limited number of inclusions in each sample and is approximately –21°C, which is characteristic for the H₂O-NaCl system. Also due to the small size of the inclusions, only
a limited number of final melting and homogenisation temperatures could be measured. The ice melting temperatures ($T_{m_{\ ice}}$) of fluid inclusions in the BOMZ vary between –11.8° and –8.4°C (average value of –11°C), while $T_{m_{\ ice}}$ values in RSF and RSC vary respectively between –14.8° and –5.4°C (average value of –9°C) and between –11.3° and –6.2°C (average value of –9°C). The inclusions in BOMZ homogenise between 127° and 192°C (average value of 162°C), while $T_{h_{\ Tot}}$ for the RSF and RSC fall between 105° and 188°C (average value of 149°C) and between 80° and 168°C respectively (average value of 140°C). The salinity and density have been calculated with the programme FLINCOR of Brown (1989), using the equation of state (EOS) of Brown and Lamb (1989). For BOMZ, the salinity calculated varies between 12.2 and 15.2 eq. wt% NaCl and the density is between 0.94 and 1.05 g/cm$^3$. The salinity and density calculated for the RSF vary between 8.4 and 18.4 eq. wt% NaCl and between 0.95 and 1.08 g/cm$^3$ respectively, while for the RSC the values are between 9.4 and 15.3 eq. wt% NaCl and between 0.97 and 1.04 g/cm$^3$.

The microthermometric data indicates that the fluids found in the authigenic quartz in the lower orebody (RSF), the upper orebody (BOMZ) and the low-grade barren zone (RSC) have a similar H$_2$O-NaCl composition and spread in homogenisation temperature and salinity. Therefore, a single fluid is proposed for the formation of the main stage of stratiform mineralisation.

**Luiswishi**

Results of the microthermometric investigation of samples of Luiswishi have been reported in El Desoucky et al. (2007). At Luiswishi, microthermometric analyses were performed on authigenic quartz in mm- and cm-sized nodules of type II and on quartz occurring in brecciated rocks. Primary fluid inclusions (up to 15 µm) in authigenic quartz are three phase (L+V+S) aqueous inclusions. The solid phase is mostly halite with few large inclusions that enclose small prismatic and rounded high birefringent crystals in addition to halite. The Th and Tsh values from the type II nodules are between 324 and 419 °C (n =42) and between 305 and 396 °C (n =93) °C respectively. Similar Th (341 to 439 °C; n =45) and Tsh (311 to 391 °C; n =78) values are obtained for the inclusions in the veins.

The inclusions in the quartz cementing the breccia show Th values between 359 and 427 °C (n =37) and Tsh values between 328 and 373 °C (n =69). In an H$_2$O-NaCl system, these Tsh values correspond to a salinity of 38.6 to 47 eq. wt% NaCl in the nodules, of 39 to 46.5 eq. wt% NaCl in the veins and of 40.4 to 44.6 eq. wt% NaCl in the breccia.

**Kipushi**

Data from Kipushi have been reported in Heijlen. 2007. More detailed information will be available in Heijlen et al. (submitted). The Kipushi Cu-Zn deposit formed along a late to post-tectonic vein and associated lodes occur at lithological contacts (de Magnée & Francois, 1988). Dolomite and quartz associated with the main stage of sulphide mineralization at Kipushi contain primary assemblages of multiphase fluid inclusions having (L+V+S), containing trapped rounded solid halite inclusions. Microthermometric study of these inclusions is hampered by the fact that most decrinate before Th or even Tsh. Minimum Th of such assemblages (for inclusions that homogenised before decripation) are between 229 and 324°C (n = 18), with Tsh between 149 and 360°C (n =
i.e. salinities between 29.6 and 43.3 eq. wt% NaCl. Late sphalerite, which formed as the youngest ore mineral, has L+V primary inclusions with Th between 80 and 167°C (n = 19). In these inclusions hydrohalite was always the latest phase to melt during cryometric runs, at metastable temperatures >0.1°C. This indicates salinities > 23.2 eq. wt% NaCl.

The data presented in Figure 16 provides the evolution of the fluids from early diagenetic conditions (nODULES Kamoto/Musonoi), to late burial – syn orogenic (Luiswishi) to the conditions after the Lufilian orogeny (Kipushi) (Figure 16). In a first phase, stratiform mineralisation (disseminated and concentrated in nodules and lenticular bands) formed due to the migration of a fluid with a temperature above 80 °C and a salinity between 8.4 and 18.4 eq. wt% NaCl. The similarity in Th and salinity of the fluid inclusions associated with the Cu-Co mineralization in type II nodules, veins and breccia of Luiswishi is a good indication that they formed from a similar fluid. This later mineralising/remobilisation period has probably a late diagenetic (e.g. pseudomorphs of type II nodules) to syn-orogenic (tectonic breccia) origin. The very high temperature of this fluid points to a deep basinal origin, eventually migrating upwards from the basement, or could be a hydrothermal fluid heated by the pre- or syn-orogenic magmatic activity. It is possible that this fluid was carrying new metals and/or remobilised the disseminated mineralization of the earlier phase and re-precipitated them in economic ore deposits. The major vein-type Cu-Zn deposit of Kipushi represents a late major mineralization phase in the Lufilian arc. The fluid inclusion characteristics (Th and salinity) are slightly lower than those of the syn-orogenic fluids. Late mineralization at Kipushi occurred at much lower temperatures and thus well after the Lufilian orogeny, as is confirmed by recent dating of Schneider & Melcher (2007).
If these results are compared to literature, it seems that the salinity and Th\textsubscript{Tot} values of Greyling et al. (2005) and Annels (1989) correspond to the fluids from the Kamoto/Musonoi nodules. The data from Annels (1989) could reflect the same hydrothermal event, but the data of Greyling et al. (2005) could be interpreted to be a little older during diagenesis, based on the structural setting of the investigated veins and the gaseous composition of the fluid inclusions. All other data are difficult to position in the basin evolution. However, compared on their salinity and Th\textsubscript{Tot}, they should be interpreted as fluids that circulated during late burial to syn-orogenic conditions and even after the Lufilian orogeny (Kamona, 1993; Kampunzu, 1998; McGowan et al. 2006, Pirmolin, 1970; Richards et al., 1988; Sweeney, 1987).

### 7.4 Radiogenic isotopes

Data from the Pb isotopic composition of sulphides of different Neoproterozoic deposits in the Copperbelt is available in literature (Figure 17). Pb data has been compiled from vein type deposits (e.g. Kabwe, Kengere, Kipushi, Lombe; Kamona et al., 1999, Walraven & Chabu, 1994) and stratiform deposits like Musoshi (Richards et al. 1988), Kinsenda (Ngoyi et al., 1993; Walraven & Chabu, 1994) and Kolwezi (Walraven & Chabu, 1994).

![Figure 17. Figure showing the 207/204Pb vs 206/204Pb ratio of sulphides from different mineralisation in the Copperbelt. The cluster of the vein-type deposits is indicated by a red circle (source data see text).](image)

Pb from sulphides of the vein-type deposits is very homogeneous and plots near the orogenic/upper crustal curve. This demonstrates that all these Zn-Pb-Cu mineralisation could have a similar origin. Furthermore, such homogeneous Pb indicates that Pb from different sources (sandstones, shales, carbonates, basement rocks, etc) is thoroughly
mixed. From the point of fluid flow mechanisms, this requires a large (extensive) hydrothermal system.

In comparison, the Pb isotopic composition from the stratiform deposits is very heterogeneous and more radiogenic (much more $^{206}\text{Pb}$, $^{207}\text{Pb}$ then expected from the model reservoirs). The Pb isotopic composition of the sulphides from Musoshi form a linear trend in the $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{207}\text{Pb}/^{204}\text{Pb}$ diagram (Richards et al., 1988). The heterogeneous, radiogenic and linear Pb-isotopic composition of both mineralisation has been explained as the result of lateral flow of fluids, scavenging Pb from their laterally continuous aquifers.

8. Metallogenic model for stratiform mineralisation in the Copperbelt

8.1 Cu-Co mineralisation and satellite imagery

A systematic study of the occurrence of the Cu-Co mineralisation in the Katanga Copperbelt, combined with satellite imagery, shows that the mineral occurrences are mainly stratiform and limited to the R-2 group (142 of 169 stratiform deposits). The stratigraphic control on the mineralisation has already been widely described in literature (e.g. Cailteux et al., 1994; Demesmaecker et al., 1962; François, 1973; Oosterbosch, 1962; Unrug, 1988), i.e. mineralisation occurs mainly in the Kamoto Dolomite and the Dolomitic Shales. These formations can often be easily identified by satellite imagery as they occur as ridges due to their resistance to weathering. These ridges, consisting of R-2 rocks, occur along the margins of thrust sheets and along fault zones in these thrust sheets. The interpretation of satellite images and data on the mineralisation indicates that mineralisation outside the R-2 formed dominantly by enrichment.

The majority of the described economic ore deposits are related to the presence of a lineament (Table 2). This is not only the case for the enrichment mineralisation, but also for the economic stratiform ore deposits belonging to the R-2 sub-group. Economic stratiform Cu-Co mineralisation are situated at the intersection of the Roan strata and a crosscutting lineament. The latter could reflect a fracture zone or even faults that formed the migration pathway for descending waters that caused the supergene enrichment and possibly for ascending mineralising fluids that formed the primary mineralisation. Raybould (1978) stressed that the stratiform copper mineralisation of the southern part of the Copperbelt are located along a northwest-southeast-trending lineament along which fluids migrated upwards from the basement to form the copper mineralisation.

8.2 Early diagenetic origin

The first sulphides precipitated in the stratiform mineralisation are framboïdal and euhedral pyrite (and idiomorphic chalcopyrite) (Figure 18), which pre-date the cobalt and the main copper mineralisation (e.g. Oosterbosch, 1951; Bartholomé, 1974; Bartholomé et al., 1971; 1972; Cailteux, 1994). Framboïdal pyrite is characteristic for an early diagenetic precipitation. This sulphide –and possibly the chalcopyrite– precipitated at temperatures between 20 and 60°C and at near-neutral pH due to bacteriogenic processes (cf. Bartholomé, 1974; Fleisher et al., 1976; Sweeney et al., 1986; 1991; Okitaudji, 1992; 2001). This first mineralisation forms a minor part, never exceeding 1%.
A purely syn-sedimentary origin of the stratiform mineralisation is abandoned based on the lack of systematic correlation between transgressive/regressive events and sulphide zonation and based on the discontinuity of the mineralisation within a single lithostratigraphic unit (Annels, 1974; Sweeney & Binda, 1994; Cailteux et al., 2007).

8.3 Diagenetic origin

The second, main generation of sulphides replaces or includes the sulphides of the first generation (Bartholomé, 1974). The association of the cobalt mineralisation and a major part of the copper mineralisation with diagenetic minerals indicate that they precipitated during further diagenesis (Figure 19; cf. Bartholomé, 1974; Bartholomé et al., 1971; 1972; Cailteux, in press; Dimanche, 1974; Unrug, 1988). This generation formed parallel to the bedding, occurs in lenses or nodules. The sulphide mineralisation consists of carrolite, bornite, chalcopyrite, digenite and chalcocite. Typically is the association of the sulphides with authigenic quartz, either disseminated in the host-rock or associated with pseudomorphs after anhydrite nodules. The pseudomorphs have been found in samples of all the studied deposits, from Kamoto over Kambove to Luiswishi. The authigenic quartz is followed by the precipitation of coarser grained dolomite cements, which can also contain mineralisation. Depending on the specific mineral deposit, digenite, chalcocite, chalcopyrite and carrolite have been found in this dolomite cement.

Petrographic arguments for the exact timing of the main mineralisation within the burial history are lacking. However, Sweeney et al. (1986) and Sweeney & Binda (1989; 1994) proposed that the bulk of the sulphides formed before appreciable burial of the sediments. A diagenetic origin is supported by the Re-Os dating of chalcopyrite in the Konkola deposit (Zambia) at 816 ± 62 Ma (in Selley et al. 2005). This mineralisation phase could be related to the early Katangan rifting of the basin during the Roan, leading to the formation of a passive continental margin (Muchez et al. 2007).
Microthermometric results of fluid inclusions in the authigenic quartz indicate the presence of a fluid with a minimum temperature between 80° and 195°C and salinity between 8.4 and 18.4 eq. wt% NaCl. These data demonstrate that during the main mineralisation, relatively hot saline fluids migrated through the Roan. Ore precipitation was caused by the upward migration of relatively hot fluids from the deeper subsurface into the Roan strata.

A similar temperature and salinity range (9-16 eq. wt% NaCl) is reported by Annels (1989) for fluid inclusions in early quartz associated with pyrite, chalcopyrite, pyrrhotite and bornite from the Chambishi deposit (Zambian subprovince). Annels (1989) proposed a model that includes the upward migration of the mineralising fluid along faults from the basement and a lateral flow in permeable arenites and below impermeable horizons such as the Ore Shale.

A mentioned above, the presence of pseudomorphs after anhydrite are widespread throughout the Copperbelt. It seems that these are associated with the main part of the mineralisation. A model is proposed by Muchez et al. (2008) that explains the origin of the pseudomorphs after anhydrite, taking several observations into account:

1) the replacement of the anhydrite by dolomite and the subsequent replacement of the dolomite by sulphides and quartz, preferentially along the rim of the nodule;
2) the typical association of the sulphides with large amounts of authigenic quartz;
3) the dissolution of the detrital quartz in the siltstones;
4) the negative $\delta^{34}$S values of the metal sulphides in the pseudomorphs;
5) the depletion in the oxygen and carbon isotopic composition of the dolomites compared to marine Neoproterozoic dolomites;
6) the high salinity of the mineralising fluid;
7) the variation in the amount of the minerals (dolomite, AQ and sulphides) present in the pseudomorphs;
8) the preferential occurrence of carrolite in layers with pseudomorphs.

If an average $\delta^{34}$S value of 17.5‰ V-CDT is taken for the Late Neoproterozoic seawater (Claypool et al. 1980), the calculated $\Delta_{SO_4-H_2S}$ values for the copper-cobalt sulphides in the evaporite pseudomorphs would suggest bacterial sulphate reduction (BSR) process for the origin for the sulphide (Ohmoto 1986, Machel et al. 1995). The net mass balance reaction for the sulphate reduction can be described by the following general equation (Machel 1987):

$$\text{organic matter} + \text{SO}_4^{2-} \rightarrow \text{altered organic matter} + \text{H}_2\text{S (HS)}^- + \text{HCO}_3^- (\text{CO}_2) + \text{H}_2\text{O} \ (1)$$

The reductant is provided by organic matter in the siltstones (Cailteux 1994) or by cyanobacterial mats in the laminated carbonates (Bartholomé et al. 1972). As a by-product of this BSR, calcite most often replaces calcium sulphate (Machel 2001). This is supported by the widespread occurrence of calcite cements in environments affected by BSR (Sassen et al. 1988). In addition to calcite, dolomite may form when the host carbonate rock is a dolostone (Machel et al. 1995, Machel 2001). The earliest mineral that
replaced anhydrite in the dolomitic siltstones and dolomites of the Roan group was indeed dolomite:

$$\text{Ca}^{2+} + \text{Mg}^{2+} + 2\text{CO}_2 + \text{H}_2\text{O} \Leftrightarrow \text{CaMg(CO}_3\text{)}_2 + 4\text{H}^+$$  \hspace{1cm} (2)

The saline, metal-bearing external fluids migrated through the sediments and in the neighbourhood of the pseudomorphs, the Cu, Fe and Co cations in solution reacted with the H$_2$S formed during BSR, resulting in the precipitation of the sulphides:

$$\text{Me}^{2+} + \text{H}_2\text{S} \Rightarrow \text{MeS} + 2\text{H}^+$$  \hspace{1cm} (3)

An important observation for the interpretation is the close association of sulphides with large amounts of authigenic quartz in the pseudomorphs. Firstly, quartz is a typical replacement product of the silicification of evaporites or precipitates from a sulphate-rich fluid. Secondly, precipitation of large amounts of authigenic quartz is only possible if the ambient fluid was rich in silica and sufficient amounts of such a fluid migrated through the siltstones and dolomites. Si-rich fluids are characterised by an elevated pH (Hesse 1989). Bartholomé et al. (1972) already noticed that the origin of the large amount of silica is an important geochemical problem, which is perhaps directly related to the genesis of the orebodies. The significant amount of copper-cobalt sulphides in the pseudomorphs implies an open to semi-open system for the mineralising fluids, allowing the mass transfer of the metals (Garven 1995). The saline, metal-bearing fluids migrated through the sediments and in the neighbourhood of the pseudomorphs, reacted with the H$_2$S formed during BSR, which resulted in the precipitation of the sulphides. Precipitation firstly occurred around the rim of the pseudomorphs and continued towards the core. Sometimes the core still consists of anhydrite. During precipitation of dolomite and sulphides, hydrogen ions are released (reactions 2 and 3), which are interpreted to have caused a decrease in pH and precipitation of authigenic quartz from an alkaline solution (Williams and Crerar 1985, Hesse 1988). Brown (2005) already suggested that the mineralising fluid in sediment-hosted stratiform copper deposits could be buffered by silicate constituents of the basin fill. The solubility of quartz increases significantly with pH (Williams and Crerar 1985). The decrease in pH may also cause the dissolution of earlier formed dolomite in the pseudomorphs and of the more fine-grained dolomite in the host-rock. Due to the fine-grained nature of the host-rock dolomite, this could not be observed. However, dissolution cavities in the dolomite filled with AQ and sulphides are observed. The dissolved dolomite forms an additional source for the dolomite cement that later replaces the anhydrite in the more central parts of the anhydrite nodule.

Changing physico-chemical conditions (Eh-pH, a$_{S2}$) and varying contributions of mineralising fluids (temperature, amount of metals available) cause variations in sulphide, AQ and dolomite precipitation. Although in general AQ and sulphides replace dolomite, different successive phases may be recognised and the amount of the three minerals may vary significantly within the nodules between different layers. However, within one layer the composition of the pseudomorphs is rather constant. The replacement of anhydrite nodules from the rim towards the centre is typical for these
pseudomorphs and implies that transport from and towards the centre remains possible (Ulmer-Scholle et al. 1993; Alonso-Zara et al. 2002).

A question which still has to be answered is the source of silica in solution. The dissolution of detrital quartz has been observed in the dolomitic siltstones, resulting in the absence of detrital quartz in layers characterised by clusters of authigenic quartz and by the formation of massive authigenic quartz with randomly oriented phyllosilicates enclosed. This dissolved silica may have formed the source of the authigenic quartz, however, other sources should not be excluded.

Characteristic is the high amount of carrolite in rocks rich in pseudomorphs after anhydrite. This can be explained by the high sulphur activity in the host-rock due to BSR. Craig et al. (1979) demonstrated that the mineral that precipitates in the Cu-Co-S system is mainly dependant of the temperature and the sulphur activity. Carrolite especially forms at very high sulphur activities, which is in agreement with the large availability of sulphur in the anhydrite nodules.

The source of high-salinity base metal mineralising brines has been investigated in numerous studies. Mostly, the brines were generated by evaporation of sea water in a restricted environment. The high salinity fluids migrated in the subsurface and often into the underlying basement (Selley et al. 2005), where they obtained a high temperature. The downward, often deep circulation of saline fluids results in a widespread dolomitisation (e.g. Jones et al. 2003). Also in the Copperbelt, evaporitic conditions existed during the sedimentation of the Mines and Dipeta Group and evaporative brines formed as evidenced by the abundance of pseudomorphs after anhydrite and early diagenetic magnesite (Cailteux et al. 2005).

The carbon and oxygen isotopic composition of the host rock dolomites partly falls within the range of Neoproterozoic marine dolomites (Veizer and Hoefs 1976; Lindsay et al. 2005). Such a marine origin is in agreement with the dolomitising models proposed by Bartholomé et al. (1972) and Cailteux (1977, 1994). Bartholomé et al. (1972) envisaged dolomitisation by the reflux of hypersaline fluids that formed in the lagoonal sedimentary environment. Dolomite precipitation in an intertidal to supratidal environment was proposed for the fine-grained dolomites by Cailteux (1974, 1994). Dolomites that form in a marine environment during sedimentation and early diagenesis are often partly or completely recrystallised (Land 1980; Nielsen et al. 1994; Smith and Dorobek 1993). The carbon and oxygen isotopic values of the dolomites that are below the marine isotope values are explained by recrystallisation, possibly under the influence of the mineralising fluids that migrated through the rocks.

The isotopic composition of the dolomites replacing the anhydrite nodules is also lower than the isotopic composition of marine dolomites. The lowest $\delta^{18}O$ value of -11.5‰ V-PDB can be explained by dolomite precipitation from seawater at a maximum temperature between 55 and 70°C (cfr O’Neil et al. 1969), and thus within the temperature range of bacterial sulphate reduction. Dolomites with higher $\delta^{18}O$ values precipitated at lower temperatures. The low $\delta^{13}C$ values are the result of preferential
incorporation of $^{12}$C, generated during the oxidation of organic matter (Irwin et al. 1977; Sweeney and Binda 1989). This oxidation can immediately be related to reaction (1).

8.4 Late diagenetic and syn- to post-orogenic mineralisation

The mineralised sediments are commonly cut by fissures, which are attributed to tectonic deformation (Figure 20). In Congo, the orebodies were tectonically dismembered during the Lufilian orogeny and form part of thrust sheets (e.g. Demesmaecker, 1962; François, 1974) related to the Lufilian orogeny.

Figure 20. Schematic reconstruction of the remobilisation of existing mineralisation and precipitation of mineralisation in veins during late-burial syn-orogenic conditions. The mineralisation formed during late diagenesis, syn-to-late orogeny are indicated by the yellow colour.

In Musonoi and Kamoto, the fractures are filled with the dolomite generations III and IV and chalcocite-digenite-hematite. The deposits in the central and eastern part of the Copperbelt show numerous veins crosscutting the mineralised beds. Numerous authors (e.g. Bartholomé et al., 1972; Bartholomé, 1974; Lefebvre, 1976; Cailteux et al., 1994; Dejonghe, 1997; Garlick, 1962; Hoy and Ohmoto, 1989; Maree, 1963; Winfield and Robinson, 1963) identified mineralising phases associated with fractures and fissures.

Sulphur isotope analysis have been carried out on Cu-Co sulphides from recrystallized lenses, nodules, veins and breccia cement at Luiswishi (El Desoucky et al in press). As mentioned above, the reported $\delta^{34}$S values are similar as for the sulphides associated with the original type I nodules and lenses. The explanation of this similarity is speculative. Possible interpretations include an in-situ remobilization of the first phase Cu-Co sulfides during the second fluid migration phase (cf Lerouge et al., 2005) or mixing between sulfur from thermochemical sulfate reduction (TSR) of evaporite relicts and sulfur from the first phase Cu-Co sulfides (El Desoucky et al in press).

Fluid inclusions have been investigated in type II nodules, veins and breccia at Luiswishi (El Desoucky et al. 207). These have been interpreted to have likely a late diagenetic (e.g. pseudomorphs of type II nodules) to syn-orogenic (tectonic breccia) origin. The similarity in Th and salinity of the fluid inclusions favours a similar fluid. The very high temperature of this fluid points to a deep basinal origin, eventually migrating upwards from the basement, or could be a hydrothermal fluid heated by the pre- or syn-orogenic
magmatic activity. It is possible that this fluid was carrying new metals and/or remobilised the disseminated mineralization of the earlier phase and re-precipitated them in economic ore deposits. The major vein-type Cu-Zn deposit of Kipushi, which has been dated to be post-orogenic Schneider & Melcher (2007), represents another major mineralization phase in the Lufilian arc. The fluid inclusion characteristics (Th and salinity) are slightly lower than those of the syn-orogenic fluids. Late mineralization at Kipushi occurred at much lower temperatures and thus well after the Lufilian orogeny.

Temperatures up to 400°C are recorded from fluid inclusions in post-ore vein quartz from Musoshi (Richards et al., 1988) and Chambishi (Annels, 1989). These high temperatures have been interpreted as the maximum burial and metamorphic temperatures reached in the area (Richards, et al., 1988) or a possible magmatic event (Annels, 1989). Richards et al. (1988) obtained a U/Pb rutile age of 514 Ma for this late hydrothermal activity, which they related to compressional deformation and metamorphism associated with the Lufilian orogeny.

Tectonic deformation and metamorphism during the Lufilian orogeny, although resulting in a set of cross-cutting (high-temperature) mineralised veins, seems -in general- not to have resulted in significant remobilisation or redistribution of the sulphides (Hoy and Ohmoto, 1989; Sweeney et al., 1991; Sweeney and Binda, 1994; Winfield and Robinson, 1963). Minor remobilisation of stratiform ores is shown by few crosscutting mineralised veins surrounded by centimetre wide zones within which stratiform sulphides have been depleted (Cailteux et al. 2007). However, the observation that economic grades of mineralisation in Roan rocks occur along lineaments, suggests the importance of the fractures as pathways to control the circulation of mineralising fluids.

8.5 Supergene enrichment
The ore deposits became weathered during their exposure, which resulted in the formation of supergene enriched ore bodies. These deposits dominantly formed in Roan rocks, but are also identified in Nguba and Kundelungu rocks. The intense meteoric alteration not only formed numerous minerals, but the process also resulted in a major increase of the copper content (from a few percent up to 25% Cu), which is of great economic importance (e.g. Demesmaeker et al., 1962; Garlick, 1962; Horscroft, 1963; Lefebvre, 1974; Oosterbosch, 1962; Ralston, 1963).

![Figure 21. Schematic reconstruction of the supergene enrichment of the existing primary sulphide mineralisation at the surface. The deposits formed by supergene enrichment are indicated by the green spots.](image-url)
This discussion indicates that, although the main phase of the mineralisation in the Katangan Copperbelt in particular and more generally in the Central African Copperbelt occurred during diagenesis in a general sense, sulphide mineralisation/remobilisation appears more specifically to have extended over a long period of diagenesis beginning with early diagenesis and continuing during burial and tectonic deformation. Supergene enrichment followed in the near-surface environment. The relation between the occurrence of mineralisation and the lineaments indicates that these lineaments exerted an important control on the migration of the fluids. A multiphase origin of the stratiform Cu-Co mineralisation is also proposed by Cailteux et al. (2007). These authors conclude that a first group of sulphides grew at temperatures less than 100°C, while a second group of mineralisation formed during the Lufilian orogenesis from metamorphic fluids, due to the reworking of syn-genetic to diagenetic mineralisation. Uranium mineral occurrences in the Lufilian fold-and-thrust belt occur along strike slip faults and are interpreted to represent hydrothermal remobilisation of previously existing diagenetic uranium mineralisation (Cailteux, 1997; Loris, 1996; Meneghel, 1981).

9. Conclusion

A multiphase origin of the stratiform ores in the Lufilian fold-and-thrust belt in the Democratic Republic of Congo is proposed based on literature and based on new data from an archive compilation, satellite imagery interpretation, petrography and fluid inclusion research. 230 mineral occurrences from the Katanga Copperbelt in the Democratic Republic of Congo (DRC) are reviewed. The majority of the Cu-Co occurrences is stratiform and occurs in R-2 sediments. On satellite images, these mineral occurrences can be located on ridges, which occur near faults or along the margins of thrust sheets. Mineral occurrences outside the R-2 have mainly formed by enrichment due to remobilisation in the supergene environment.

The combination of satellite imagery and data on the mineralisation shows that economic mineralisation – stratiform and enrichment – can be related to lineaments. Economic stratiform occurrences have been identified at the intersection of the R-2 and a crosscutting lineament. The latter could reflect a fracture zone or even a fault that formed the migration pathway for descending waters that caused the supergene enrichment or possibly for ascending mineralising fluids to form primary stratiform mineralisation.

The stratiform Cu-Co occurrences have often been attributed to a strictly syn-sedimentary, early diagenetic, burial diagenetic or syn-orogenic origin. During this research, it is postulated that mineralisation formed during numerous successive mineralising stages. The main phase of the mineralisation in the Central African Copperbelt occurred during diagenesis, well before the Lufilian orogeny. However, mineralisation started already during early diagenesis and continued long-time after the orogeny in the near-surface environment. After deposition of framboidal pyrite during early diagenesis, Cu and Co–rich fluids circulated through the subsurface and formed the main mineralisation during diagenesis before folding and faulting of the rocks. The microthermometric data of the authigenic quartz demonstrates that ore precipitation was
caused by the upward migration of relatively hot fluids from the deeper subsurface into the Roan strata (DRC) or it formed at depth during burial from basinal brines that could have originated from the Roan or even basement strata (Zambia). The Lufilian orogeny caused numerous fractures and faults cutting the mineralised beds. Circulation of fluids along these faults resulted in additional remobilisation phases and important supergene enrichment.

The host rock of the stratiform Cu-Co ore deposits in the Mines Subgroup of the Roan Group has been recrystallized and dolomitised by the infiltration of hypersaline fluids that formed during the lagoonal evaporitic conditions during the sedimentation of the Mines and Dipeta Group, as is indicated by the oxygen isotopes that partly fall within the range of Neoproterozoic marine dolomites. This hypersaline fluid is interpreted to have migrated further downwards into the basement to get likely charged with metals.

Pseudomorphs of anhydrite nodules and lenses are abundant in the Mines Subgroup. These pseudomorphs consist of dolomite that has been replaced by sulphides and authigenic quartz. The replacement of the nodules and lenses is interpreted to be related to the bacterial reduction of the sulphate (BSR) of the original anhydrite, based on the negative $\delta^{34}S$ values ($\delta^{34}S = -10.3 – 3.1%o$ V-CDT) of the sulphides in the evaporite pseudomorphs. The generation of CO$_2$ during this reaction caused firstly the precipitation of dolomite. The distinct negative carbon isotopic composition of the dolomite in these pseudomorphs (-7.10 and -9.93‰ V-PDB) are due to the incorporation of light carbon, interpreted to be generated by the oxidation of organic matter during the BSR.

When the metal-bearing fluid reached the site of the BSR, it reacted with the BSR-derived H$_2$S at the site of the pseudomorphoses and precipitated the Cu-Co sulphides. The H$_2$S reacted with e.g. Cu, Co and Fe to form sulphides such as carrolite, bornite, chalcopyrite and chalcocite. During dolomite formation, BSR and sulphide precipitation, hydrogen ions are released, causing a decrease in pH and the precipitation of quartz, post-dating dolomitisation and often also the sulphides. The mineralising fluid had an intermediate salinity (8 to 18 eq. wt% NaCl) and is thought to represent evaporated seawater of Roan age.

A second mineralisation and/or remobilisation phase is characterised by the occurrence of Cu and Co sulphides in dolomite and quartz veins that crosscut the nodules and lenses of the first main mineralisation phase. A remobilisation is suggested by the identical mineralogy of the sulphides of the first phase and in the veins and geochemical similarities between both phases. Higher salinity and higher temperature fluids are responsible for this phase. Different periods can tentatively be suggested for formation of these mineralised veins. The veins could have formed during late diagenesis (cfr Selley et al. 20005). Secondly during the Late Roan, when volcanic and magmatic rocks were emplaced in the continental rift setting and when an elevated heat flow was present. Thirdly during the Lufilian orogeny between 592 and 512 Ma (Rainaud et al. 2005) that caused the deformation of the Katangan sediments. Radiometric dating of sulphides at Nkana, Chibuluma and Nchanga (Zambia) revealed mineralisation ages around 583 and 526 Ma (Barra et al. 2004). Vein-type polymetallic mineralisation also occurs in the
Copperbelt (e.g. Kipushi, DRC). Structural and radiometric evidence indicates these deposits formed during and after the Lufilian orogen.

The ore deposits became weathered during their exposure, which resulted in the formation of secondary supergene enriched ore bodies, which can be followed until a depth of several hundreds of meters.

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