



ELSEVIER

Ore Geology Reviews 9 (1994) 105–130

ORE GEOLOGY
REVIEWS

Metallogeny of the northeastern Kibara belt, Central Africa— Recent perspectives

W. Pohl

Institute of Geosciences, Technical University, P.O. Box 3329, D-38023 Braunschweig, Germany

(Received April 13, 1993; revised version accepted October 20, 1993)

Abstract

The Kibara belt is an intracontinental mobile belt formed between 1400 and 900 Ma within a craton of Lower Proterozoic age. The belt's evolution started by early rifting at about 1400 Ma, and continued by transition into a marine basin filled by clastic sediments > 10 km thick, with minor basic and acidic volcanic rocks. At about 1300 Ma, the pile was deformed by thrusting and folding of the main Kibaran Orogeny and intruded by numerous large syn-orogenic granite bodies. Wide areas of the Kibaran experienced only very low-grade metamorphism, but halos of high-temperature/low-pressure metamorphism surround intrusive bodies. A post-orogenic phase of rifting followed at about 1275 Ma and was accompanied by the intrusion of alkaline granites and layered mafic intrusions with Ni, Cu, Co, Ti, V and possibly also PGM mineralization. Also at about this time, molasse sedimentation set in, both within the belt and in foreland troughs.

The western molasse and the whole belt were again orogenically deformed at around 950 Ma by the Lomamian Orogeny; concurrently, small metallogenetically specialized granite bodies were emplaced. Fertile members ("tin granites") of the latter produced numerous pegmatitic and quartz vein deposits with Sn, W, Nb/Ta, Li, Be and Au. Other auriferous and locally Pt-bearing quartz veins and breccias are not spatially related to tin granites, but appear near large fault zones that may be associated with deep structures separating basement blocks. Both the tin granites, and the fluids which formed gold ores not related to granites, are suggested to have been derived from metamorphic devolatilization and partial melting of older (Archean and Lower Proterozoic) basement underneath the Kibaran Orogen.

1. Introduction

The Mesoproterozoic (Plumb, 1991) Kibara belt extends from the Zambia/Angola/Zaire border in the southwest, through Zaire, Burundi, Tanzania and Rwanda to southwestern Uganda in the NNE. Its overall length is nearly 1500 km, and it attains widths of up to 400 km (Fig. 1). Economic mineralization is known mainly from its northern sectors, within three geographically separated areas, comprising the Shaba and Kivu

provinces west of the Central African Rift and a northeastern province covering Burundi, Rwanda, the West Lake district of Tanzania and southwest Uganda. The northeastern province is the main subject of this paper (Fig. 2).

Since the discovery of cassiterite deposits after the First World War, the Kibara belt has become a source of tin, and also tungsten, gold, beryllium, columbo-tantalite and lithium ores. In addition, minor exploitation of bastnaesite (REE), uranium ores, mica and semi-precious stones has

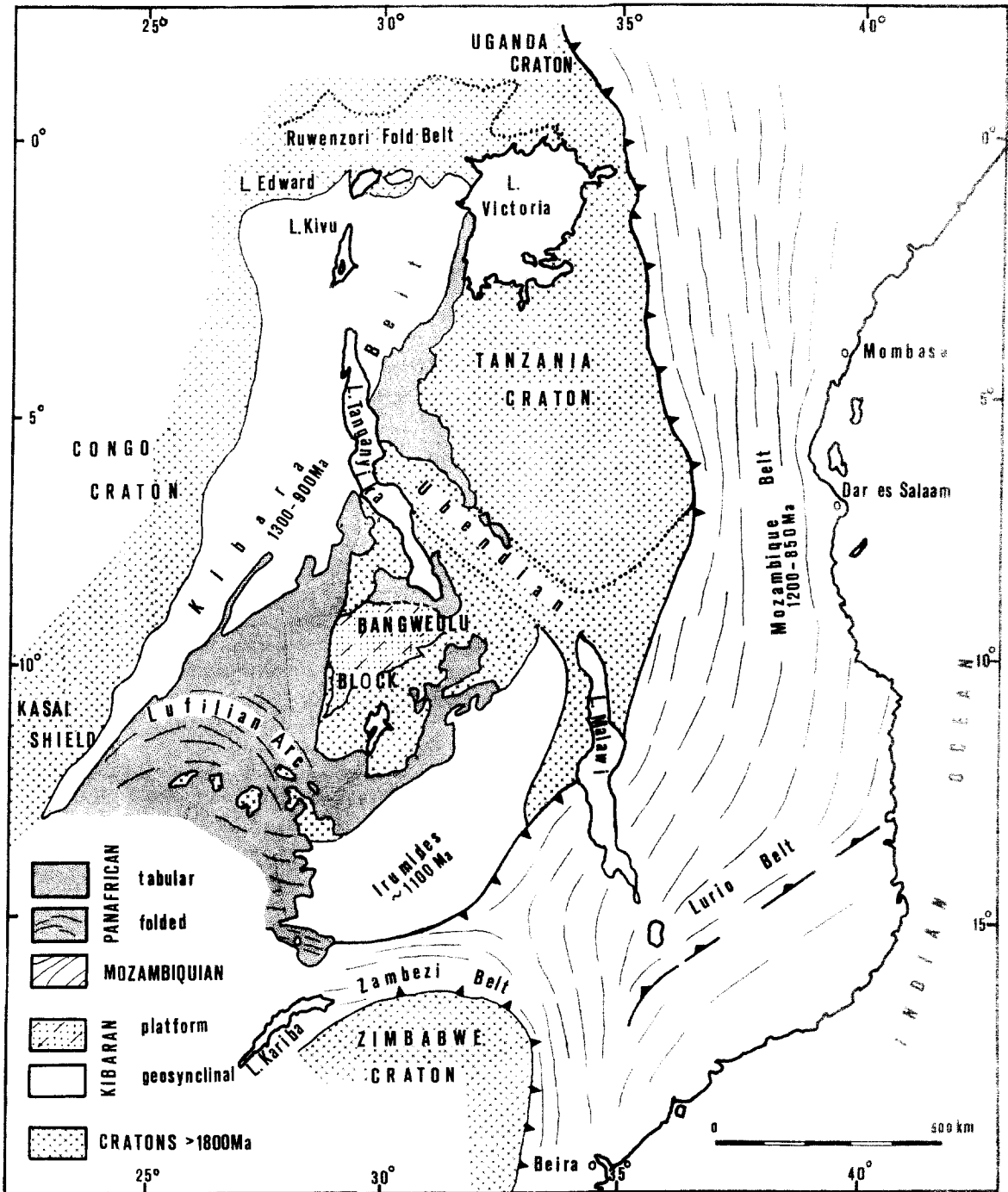


Fig. 1. Structural setting of the Kibara belt in Central Africa (after Cahen et al., 1984, with additions).

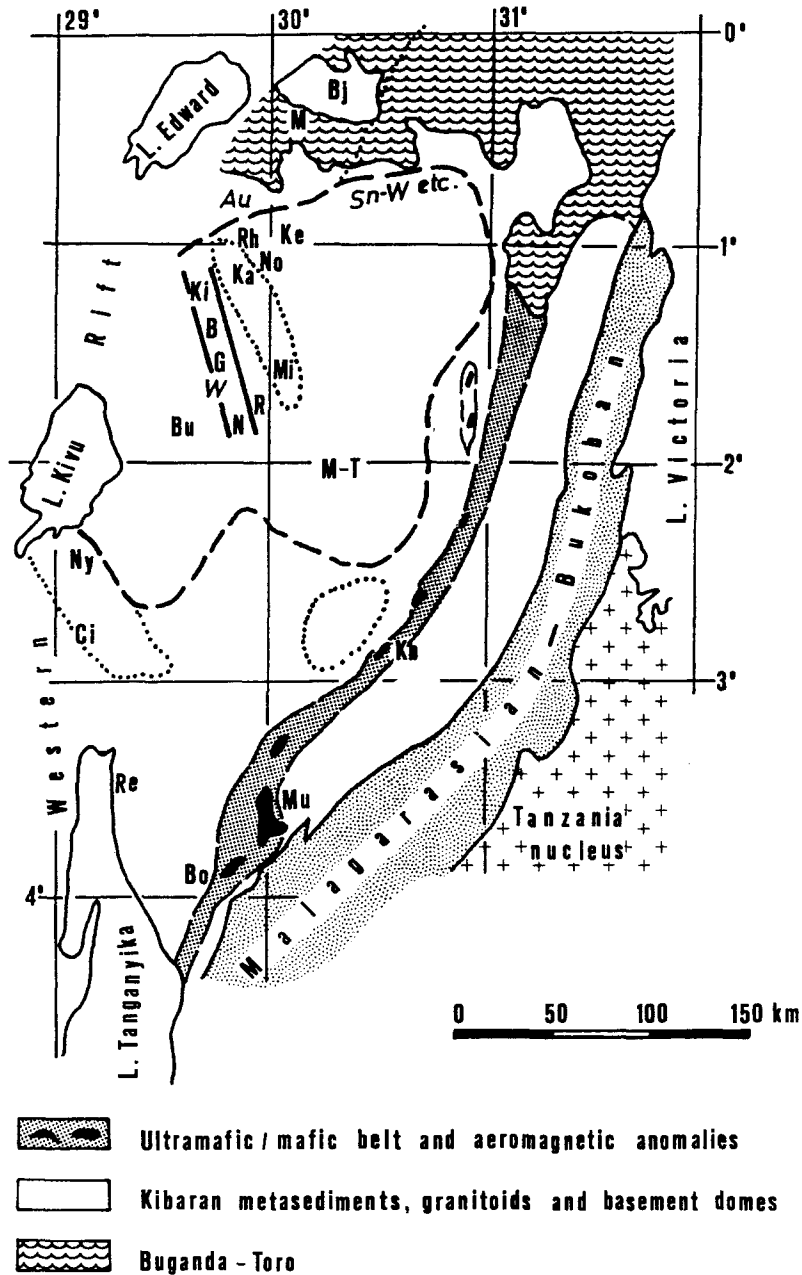


Fig. 2. Sketch map of the geology and metalliferous zones of the northeastern Kibara belt (Pohl, 1987). Mines, mining districts and localities mentioned in the text: Uganda: *Bj*—Buhweju plateau; *Ka*—Kamena Fe; *Ki*—Kirwa W; *Ke*—Karengé G4-granite; *M*—Mashonga Au; *No*—Nyamulilo W; *Rh*—Ruhiza W. Rwanda: *B*—Bugarama W; *Bu*—Buranga pegmatite; *Mi*—Miyove Au; *G*—Gifurwe W; *N*—Nyamulilo (Shyorongi) W; *Ny*—Nyungwe Au; *R*—Rutongo Sn; *M-T*—Musha-Ntunga Sn. Burundi: *Ci*—Cibitoke/Mabayi Au; *Bo*—Buhoro Ti, V; *Mu*—Musongati Ni; *Re*—Karonge REE. Tanzania: *Kb*—Kabanga Ni-Cu-Co. /W/ signifies the “tungsten belt”; dotted lines enclose auriferous zones.

taken place. On a worldwide scale, however, the belt has not been a major producer of any of these commodities. Cumulative cassiterite production in the northeastern province, for example, may be estimated at approximately 150,000 m.t., which is roughly equivalent to the maximum annual output in Malaysia. Nevertheless, mineral production has been an important economic factor for the countries concerned. At present, little mining at industrial scale takes place in the region for various reasons, although potential reserves of “traditional Kibaran commodities” are known and resources have not been fully evaluated. Increasingly, however, the belt is being explored for metals associated with layered mafic intrusions (Ni, Cu, Co, Ti, V and PGM), and its potential for various industrial minerals (andalusite, kaolin, muscovite, quartz, talc, wollastonite) is being recognized.

2. Geological setting and evolution of the Kibara belt

The Kibara belt is the westernmost of three roughly synchronous and parallel orogenic belts in central and eastern Africa, the other two being the Irumides in Zambia (Daly, 1986) and the Lurio belt in Mozambique (Jourde and Vialette, 1980; Sacchi et al., 1984; Costa et al., 1992). In the west, the Kibara belt is limited by the Congo (Zaire) craton (although the border area is not well exposed because of a younger cover), and to the east its boundary with the Tanzania craton is largely masked by a blanket of post-Kibaran platform sediments (the Bukoban and Malagarasian Supergroups). Farther south, folded Katanagan rocks (of the Lufilian arc) occur to the east of the Kibara belt. Within the belt, early Proterozoic schists and gneisses (generally of Eburnean/Ubendian age) and, less frequently, Archean granitoids, granulites and greenstones emerge in cores of antiforms and uplifted blocks. This relationship between the Kibaran rocks and their basement is exceptionally well exposed at its northern end in southwestern Uganda, where the lower Proterozoic Buganda–Toro system occurs underneath Kibaran sediments with a clear

unconformity often marked by conglomerates (Fig. 3: Pohl and Hadoto, 1990). This is a unique area, since a virtually continuous cross-section of basement and folded Kibaran sediments is accessible here; geodynamic models of the Kibaran, therefore, must account for this situation.

Although parts of the Kibara belt appear to show sub-parallel trends to older basement structures, it clearly cuts across the general structural grain of the previously stabilized craton. Accordingly, the belt cannot be considered as the remobilization of an earlier zone of structural weakness, but it is a wholly new, major Mesoproterozoic feature largely independent of the previous geological evolution.

Worldwide, the older middle Proterozoic (about 1800–1400 Ma) is characterized by continental rifting accompanied by the formation of some giant ore deposits, including Mt. Isa (Cu, Zn, Pb, Ag), Broken Hill (Zn, Pb, Ag) and Olympic Dam (Cu, U, Au) in Australia, to name but a few examples. Comparable base-metal deposits, however, have not been found until now in the Kibara belt, whose evolution essentially post-dates this major phase of rifting. Also, its characteristic mineralization is totally different, being comprised of the typical granitophilic elements Sn, W, Li, Be and Ta, with some gold. It is significant, that Sn and related mineralization first appears in large numbers of occurrences (although not in quantities of metal mined) during the Mesoproterozoic. Most of these occurrences, however, are found in anorogenic settings; the largest deposits being those of northwestern Brazil (Tassinari et al., 1989). Smaller occurrences are known from Greenland, Scandinavia, India, Congo and Australia.

Towards the later middle Proterozoic (ca. 1400–900 Ma), the predominantly tensional geodynamic regime characterizing the evolution of the continents from about 1800 to 1400 Ma changes into a compressional, and often collisional pattern, although still marked by extensional phases. Examples of collision belts are the Grenvillian belts of North America and western South America, the Albany–Fraser Orogen in southwestern Australia, the Namaqua–Natal belt in Southern Africa (Jacobs et al., 1993; Thomas

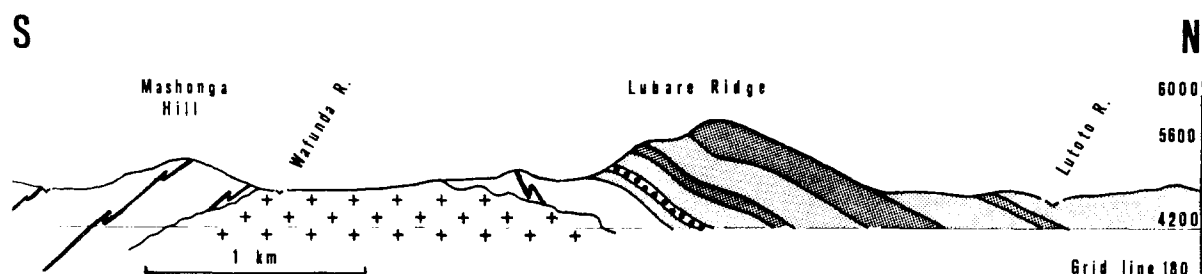


Fig. 3. Geological cross-section of the Mashonga gold field at the southern escarpment of the Buhweju Plateau, Uganda (Pohl and Hadoto, 1990). G4 granite intrude Buganda Toro amphibolite facies schists with quartzite bands. Kibaran sediments (conglomerate, quartzite, pelites) are clearly discordant. Auriferous quartz veinlets occur in all units shown, gold placers occur in the Wafunda River basin.

et al., 1994, this issue), and its equivalent in Dronning Maud Land, Antarctica. A strong, although somewhat enigmatic Kibaran imprint is also known in western Africa. Collision of late Kibaran age is further recorded from the southern Mozambique belt (Daly, 1986; Pinna et al., 1993: 950–800 Ma). Towards the North, however, Mozambiquian collision is younging (in Kenya at about 820 Ma: Key et al., 1989), and in the Arabian–Nubian shield subduction in an oceanic environment and accretion of the resulting island arcs continued until suturing and collision occurred between about 715–640 Ma.

Within this geodynamic frame, the Kibaran of Central Africa undoubtedly reflects the response of the continental crust of what may be called the “pre-Kibaran African Craton” to the formation and destruction of ancient oceans around its margins. In concurrence with the concept of “far field tectonics” (Ziegler, 1987), it is further proposed that periods of major extension and collisional events around this craton are recorded within the Kibara belt. If this hypothesis is correct, the relatively well-constrained Kibaran evolution may serve in turn as a reference for tectonic activities taking place at the margins of the craton between about 1400 and 900 Ma.

In this article, the first aim is to review and to illustrate the geological evolution of the Kibara belt *sensu stricto*, as a contribution towards a synthesis of Kibaran evolution in Africa and on other continents. The second intention is to integrate known metallogenic features of the belt

into its evolutionary history, and to propose some ideas for future strategic exploration.

2.1. Kibaran basin evolution

Kibaran rocks in the northern sector of the belt have been described as “Burundian” in the francophone countries, and as “Karagwe–Ankolean” or “Muva–Ankole” in Uganda and Tanzania. The name “Kibaran” was coined in Shaba/Zaire (Robert, 1931). Everywhere in the belt, a three-fold lithostratigraphic division of the Kibara Supergroup is recognized (for a detailed discussion; see Cahen et al., 1984). Lateral and vertical facies changes are very prominent (Baudet et al., 1988), however, so that the current basin-wide lithostratigraphic correlation must be regarded as provisional. The sediments variously represent starved basins, turbiditic environments, deltas and shallow marine siliciclastic flats. Despite these uncertainties, the following succession is reasonably well established.

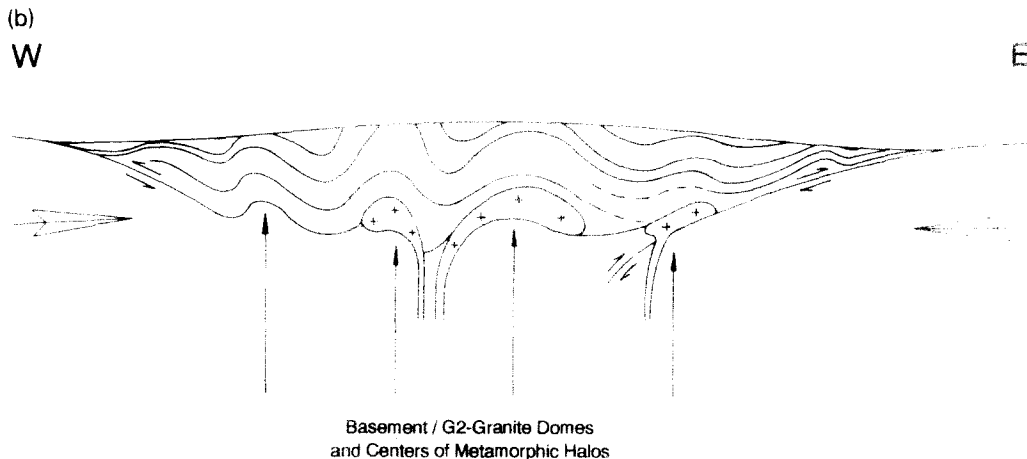
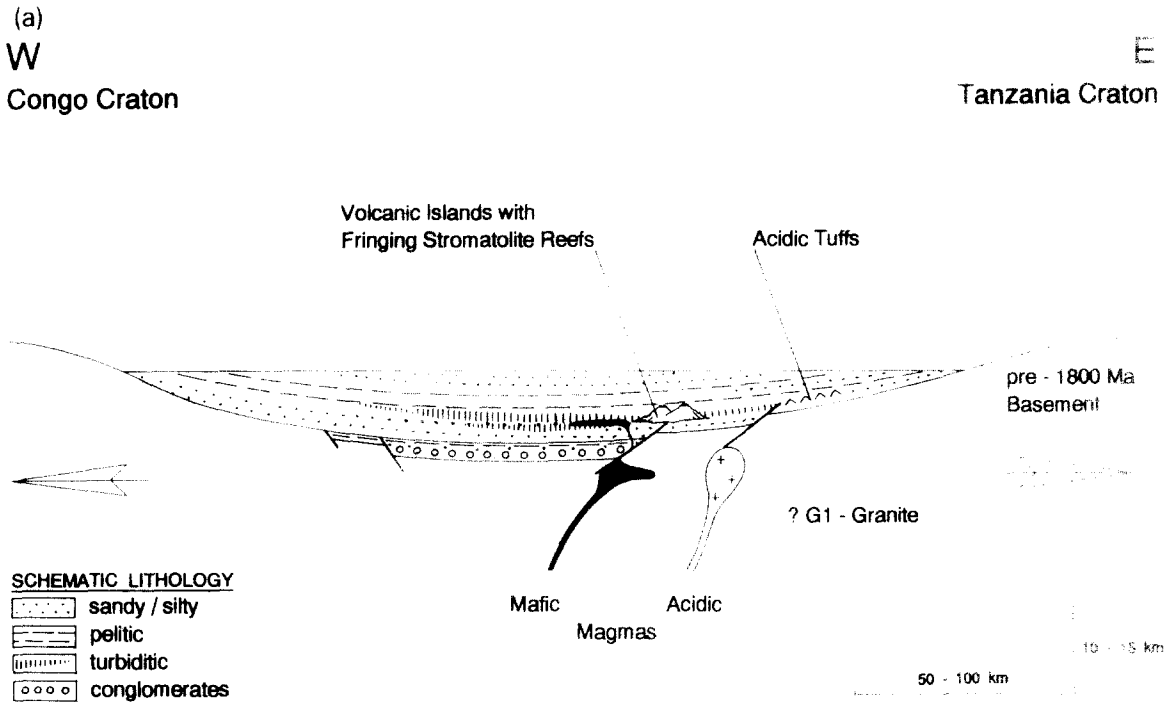
The **Lower Group** is characterized by dark, laminated pelitic sedimentary rocks with intercalations of mature quartzites, microconglomerates, sandstones and siltstones. Parts of the suite are proximal and distal turbidites. Acidic tuff bands are locally interbedded with the sediments, dated at ca. 1350 Ma in eastern Burundi (Klerkx et al., 1987). Silicified stromatolites have also been tentatively identified (Van Straaten, 1984). Fluvial conglomerates, overlying older basement, are known locally in marginal areas of the Kibara belt (S-Burundi, Uganda),

with well-rounded pebbles of quartz and of more resistant basement rocks (quartzite, pegmatite, gneiss, etc.). This indicates an early rifting event, passing into a phase of thermal subsidence, whose climax may coincide with the deposition of black shales and turbidites (mainly in the Lower and Middle Groups).

Within the belt, however, the Kibaran sedimentary rocks are underlain by their metamor-

phic equivalents or by an older metamorphic basement, or are intruded by granitoids. There, a clear basal unconformity can rarely be established.

Sedimentary structures abound in the elastic rocks (graded bedding, flute and load casts, ripples, slumping, etc.). Some of the quartzites persist over very wide areas and are useful regional marker horizons, while others are lenticular and



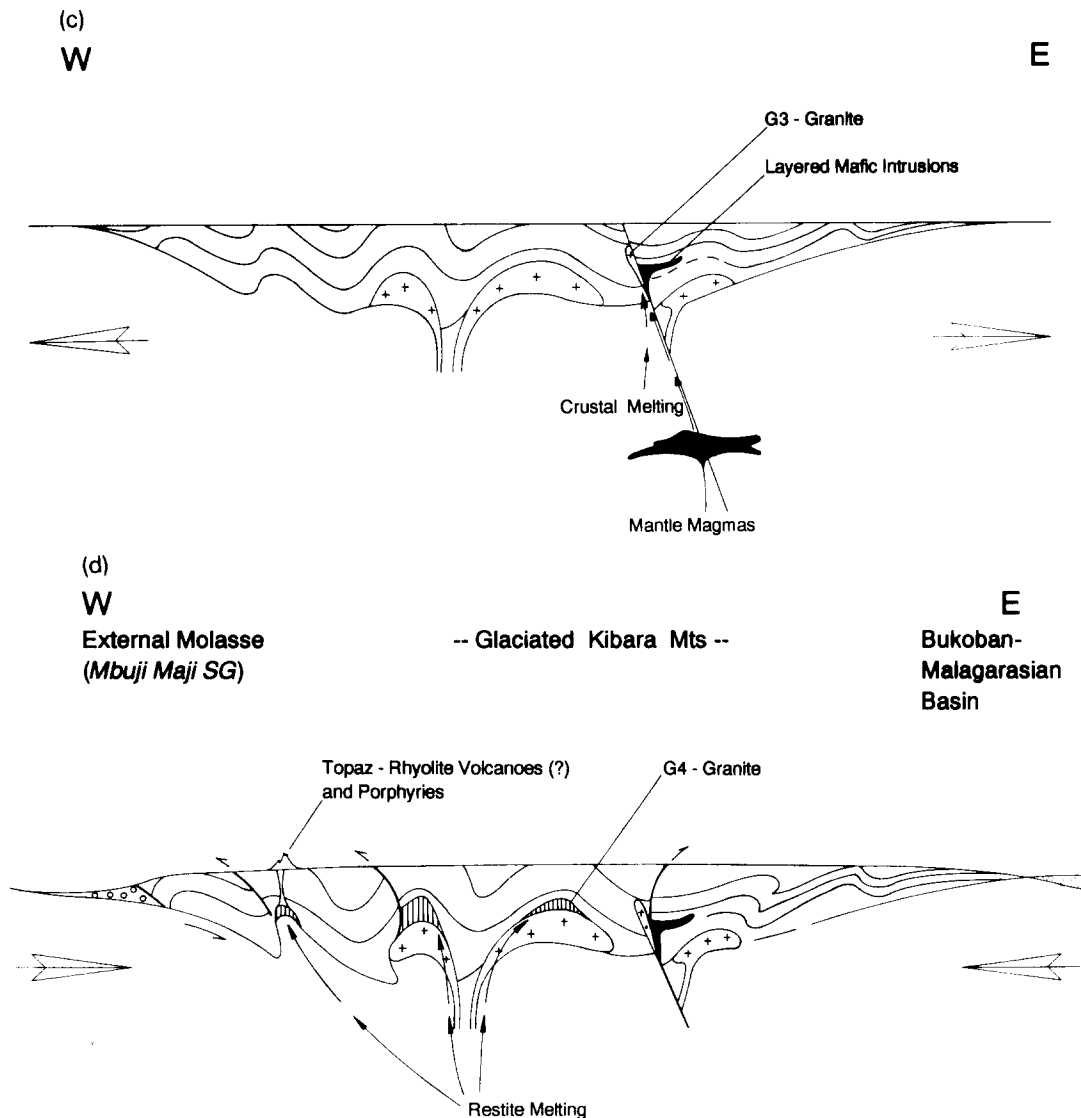


Fig. 4a. Graphical synopsis of Kibaran basin formation from about 1400 to 1300 Ma.

Fig. 4b. The Kibaran orogeny at about 1300 Ma: formation of open folds, of granite-gneiss domes and of halos of thermal metamorphism.

Fig. 4c. “Katangan” rifting affecting the Kibara belt at about 1275 Ma; mantle magmas induce crustal melting and rise along tensional faults to form layered mafic intrusions within Kibaran meta-sediments.

Fig. 4d. The Lomamian orogeny at about 950 Ma, inducing renewed compressive deformation, elevation and glaciation of the Kibara mountains, and intrusion of G4-granites. Open circles denote molasse sediments.

rather local features. Carbonates are very rare. The Lower Group attains its maximum thickness in Burundi (several 1000 m).

Clastic rocks of the **Middle Group** are clearly more arenaceous and of a lighter, often reddish colour compared with the older sequences un-

derneath. An important marker bed is taken as its base, which consists of 250 (Rwanda) to 1000 m (Burundi) of banded fine-grained and occasionally conglomeratic white to pinkish quartzite. In the west, basaltic volcanic edifices and sills have been mapped near the top of the Middle

Group, associated with thin sheets of dacitic, trachytic and rhyolitic rocks (Ntungicimpaye and Tack, 1992). Some of these volcanoes may have formed islands fringed by stromatolitic reefs which are preserved as discontinuous banded or massive carbonates, usually dolomites. Several attempts have been made to classify the tectonic setting of the basaltic rocks (Ntungicimpaye, 1984; Jung and Meyer, 1990); until now, however, with ambiguous results.

Within deeper levels of the Kibaran, granites (designated G1 or Gr 1 and 2: Klerkx et al., 1987) were intruded, apparently while sedimentation still continued (Fig. 4a). Rb/Sr dates indicate an age between 1330 and 1250 Ma for these rocks; this, however, must be considered to represent cooling ages rather than actual time of intrusion, and emplacement before 1300 Ma is indicated by geological relations (Tack et al., 1994). Most probably, these magmas represent melts produced in the lower crust underneath the belt, due to ponding of basic magma at the base of the lithosphere; magmas, whose surficial expression may be the tholeiites found in the Kibaran sedimentary pile (Klerkx et al., 1987).

The **Upper Group** occurs in major synclinal structures only. Its lowest parts consist of rather immature clastic sediments, in some areas with a slight erosional unconformity at its base marked by polymictic coarse conglomerates with pebbles of quartz and underlying rocks. Quartzites are frequently ferruginous, and reddish or grey colours predominate. Towards the top, fine-grained whitish siltstones and shales with chert laminae appear, probably indicating an environment of saline lakes (Baudet et al., 1988). Some of these basins may be Kibaran intramountainous molasse rather than part of the geosynclinal suite.

2.2. Deformation and structural evolution

At around 1300 Ma occurred the main compressional deformation (D2) of the belt (Fig. 4b). This phase is marked by granites (designated G2 or Gr3) intruding syntectonically the upright folds (F2), which are the most characteristic structural feature of the Kibara belt. The geochemical character of these granites is colli-

sional (Rumvegeri and Katabarwa, 1990). Around the granites relatively restricted halos of more intense, ductile deformation (with a schistosity S_1 parallel to bedding planes which is often attributed to a deformation phase D1) and of higher metamorphic grade appear, commonly with andalusite and staurolite in the metapelites. Elsewhere, the Kibaran sedimentary rocks are of very low metamorphic grade. Accordingly, metamorphism is generally of a high- T /low- P type typical for magmatic arcs. Metamorphic rocks indicating medium- to high- T /medium- P conditions are rare and their significance remains to be studied.

D2-folding was accompanied by the formation of an axial planar cleavage (S_2) at shallow levels; shortening across the belt was probably less than 50%. It may be assumed that this deformation caused locally (S Burundi) a decollement between basement and the overlying Kibaran sediments.

Kibaran granites of all ages intruded preferentially in anticlinal positions, and G1, G2 and G4 often form intrusive complexes associated with updoming of basement and Kibaran cover. However, only in rare cases is it possible to prove the presence of pre-Kibaran rocks in these granite-gneiss domes (Theunissen, 1989).

This stage in the evolution of the Kibara belt, the Kibaran Orogeny *sensu stricto*, formed the belt's main structural features. Considering the previous prolonged magmatic activity, the crust underneath the belt may have been still hot. Its thickening by convergence between the Zaire and the Tanzania cratons explains the syn-tectonic crustal melting, which resulted in the production of large quantities of only acidic magmas, in contrast to the earlier bimodal suites.

Due to flexural subsidence of the foreland, an external molasse basin developed after the Kibaran Orogeny along the western margin of the belt in Shaba/Zaire. The rocks (Mbuji Mayi or Bushimay Supergroup) comprise red sandstones and quartzites, conglomerates and stromatolitic carbonates. The total thickness reaches about 3000 m. There is no indication of a comparable evolution on the eastern side of the belt. This may indicate a tendency of subfluence or A-subduc-

tion of the Zaire craton eastwards underneath the Tanzania craton during the Kibaran orogeny, although no clear vergence of deformation within the belt has been derived from structural studies.

Thickened orogenic wedges are unstable, because of buoyancy and for mechanical reasons. It may be assumed, therefore, that the Kibaran mountains formed at around 1300 Ma had already been eroded to a peneplain before the next event that can be recognized. This is the formation of rifts or half-grabens at about 1275 Ma, dated by basic intrusions and alkaline granites (G3 or Gr4) intruded along these structures in Burundi (Tack and Deblond, 1990; Tack et al., 1994). Most prominent are the layered mafic intrusions within this setting (Deblond, 1990; Deblond, 1993), forming a linear belt of nearly 400 km length. Obviously, this is again an event producing bimodal magmatic rocks due to extensional tectonics at a lithospheric scale (Fig. 4c).

This interpretation is in contradiction with earlier models derived from field studies in Burundi, which considered the structures controlling the intrusion sites to be due to lateral shearing (Klerkx et al., 1987; Theunissen, 1989). Careful re-examination (Tack, 1990) of critical exposures proved only eastward directed thrusting along these structures, clearly post-dating the intrusion of the magmatic bodies. Overall, therefore, a model of rifting and rift-related magmatism followed by later tectonic inversion during the Lomamian orogeny (see below) best explains these observations.

Possibly, some of the basins described before as intra-mountainous molasse troughs with evaporitic playa sediments may have been initiated and filled at this stage. Also, the evaporitic base (with most of the copperbelt mineralization) of the Katangan Supergroup was possibly deposited at about this time. It is tempting to associate these events and to ascribe them to a Katangan rifting, or extensional phase, affecting large parts of southern Africa.

In its type area, the Mbuji Mayi Supergroup is deformed by the folding and thrusting of the Lomamian orogeny, under an east–west directed compressive stress regime. The timing of this orogeny (about 950 Ma) is roughly coincident with

the intrusion of numerous geochemically specialized or fertile “tin granites” throughout the Kibara belt. It coincides further with sedimentation breaks in the Katangan Supergroup to the southeast, and especially with the deposition of the glaciogenic Grand Conglomérat, that provides evidence for renewed uplift, erosion and glaciation of the Kibara belt.

Within the Kibara belt, the significance of this orogeny was first recognized by Pohl (1977), as structures (D3) controlling tin and tungsten mineralization and comprised of folds and thrusts of a different geometry and direction from those due to the Kibaran orogeny *sensu stricto* (D2). Both re-folding along axes sub-parallel to earlier folds as well as crossfolding at small angles are known. Together with the earlier compressional phase, shortening across the belt may, at this stage, have reached 50% (Fig. 4d). It is necessary to add to these observations those of Tack (1990) concerning the thrusting along the structures hosting alkaline granites and layered mafic intrusions in Burundi (see above). These structures were originally formed as lithospheric extensional faults, and during the Lomamian orogeny they were rotated to accommodate shortening across the belt (inversion tectonics). In southern Burundi, a conspicuous retrograde (sericitic) decollement zone between Kibaran rocks and their Archean basement also may have formed at this stage.

A shallow molasse basin developed along the eastern margin of the Kibara belt after the Lomamian Orogeny, represented by the predominantly arenaceous sediments of the Bukoban–Malagarasian Supergroups and associated tholeiitic plateau basalts.

2.3. Intrusive rocks

The numerous granite intrusions, formed over a wide time span, are a remarkable feature of the Kibara belt. Four main age groups (G1–G4) have emerged from published Rb/Sr data (Cahen et al., 1984). The first two are approximately synorogenic, whereas the latter are usually described as post-orogenic. Detailed geochronological work in Burundi led to a proposal

to differentiate the earliest granites (G1) into Gr1 and Gr2 (Klerkx et al., 1984). The student of Kibaran geology should be aware that this has resulted in two classifications of Kibaran granites (G1–G4 versus Gr1–Gr5), both used in current publications.

Porphyritic gneissose two-mica or biotite granites and adamellites (G1) with variable initial Sr-ratios (Klerkx et al., 1984; Fernandez-Alonso et al., 1986) intruded the Lower Group subconcordantly from 1350 to about 1300 Ma. Large porphyritic peraluminous adamellitic orthogneiss and granite bodies (G2) of S-type chemistry (Klerkx et al., 1984) are associated with the first, or formed separate domes at around 1300 Ma, which is considered to date the Kibaran orogeny (Lavreau, 1984; Tack et al., 1994). The source rocks of these granites were probably Lower Proterozoic (and Archean?) gneisses forming the lower crust underneath the Kibara belt at these times. A high degree of melting may be deduced from chemical data, resulting in granodioritic magmas. These intruded the sediments at 5–10 km depth representing relatively shallow levels; their magmatic evolution was strongly influenced by differentiation and by contamination with Kibaran sediments (Fernandez-Alonso et al., 1986).

Recently, the emplacement of the mafic-ultramafic intrusions of eastern Burundi (Fig. 2) was dated by the U/Pb method to have occurred at 1275 ± 10 Ma, while the associated alkaline granites (G3) yielded an age of $1249 + 8 / - 7$ Ma (Tack et al., 1994). This bimodal suite forms a belt of intrusions about 400 km long and 50 km wide. The alkaline biotite granites and syenogranites are clearly post-tectonic in relation to the main Kibaran folding. Due to Lomamian deformation, parts of the intrusions show foliation developed under greenschist facies metamorphic conditions. The mafic and ultramafic intrusions are layered and comprise peridotites, pyroxenites, norites, anorthosites and granophyres (Deblond, 1990). Aureoles of thermal metamorphism are clearly developed and prove the intrusive character of these rocks. The obvious derivation of the mafic-ultramafic suite from the mantle as well as the low initial Sr ratios of the

granites ($Sr_i = 0.7027 \pm 0.0011$; Tack et al., 1990) are compatible with a rift origin.

Most important from the point of view of this paper are the leucocratic, sub-alkaline, and strongly peraluminous G4- or “tin” granites with ages clustering at about 976 ± 13 Ma and with an average initial Sr ratio of 0.7721 (Cahen et al., 1984). They are equigranular “aplitic” or pegmatitic, often cataclastic and locally sheared, and clearly cross-cutting. Their country rocks include older granitoids and Kibaran sediments, in rare instances rocks thought to be Kibaran molasse. Locally, in Uganda, they are found in schists forming the basement below Kibaran sediments (Fig. 3; Pohl and Hadoto, 1990). Typically, G4-granites consist of quartz, microcline, albite, and muscovite (biotite) with accessory apatite, zircon and tourmaline. Their roof is often invaded by suites of pegmatites and quartz veins which host Sn, W, and Nb/Ta mineralization. Although in Rwanda these granites are clearly of crustal derivation, as shown by extremely high initial Sr-isotope ratios (Fig. 5),

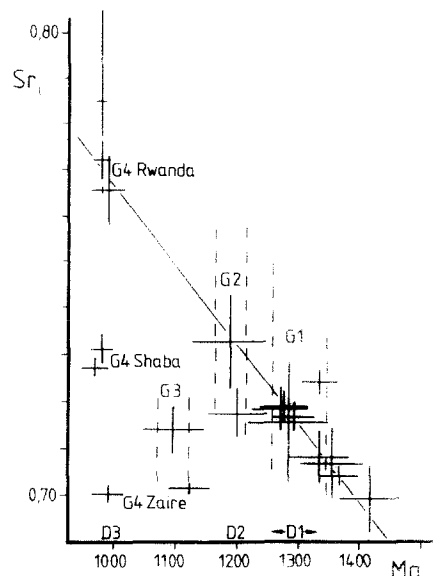


Fig. 5. Evolution of initial Sr-isotope ratios of Kibaran granites in time (Guenther, 1990). (Data from Cahen et al., 1984; Klerkx et al., 1984; Lavreau and Liégeois, 1982; Tack et al., 1990.) Note that these are Rb/Sr ages. U/Pb dating by Tack et al. (1994) brings G3 to 1250 Ma and implies a shift of G2 to > 1275 Ma.

their equivalents west of the rift in Zaire appear to be mantle-derived or strongly influenced by mantle fluids. Current diagrams for discriminating the tectonic setting of granitoids place the Kibaran G4-granites in the collisional field (Fig. 6; Guenther et al., 1989). As most of the G4-granites appear to have suffered hydrothermal alteration, this is considered as a preliminary although plausible result. Comparison with worldwide classifications of tin granites (Hutchison, 1982) remains difficult. In view of the pronounced structural control of the mineralization, and of many of the G4 intrusion sites themselves, it appears inappropriate to call them anorogenic. In addition, their shape is mostly lobate and irregular (for example, the Karenga granite, Uganda; Lowenstein, 1966), which is not at all typical of the well-defined cross-sections of subvolcanic A-type granites (Bowden and Karche, 1984). The G4-granites were probably intruded as sheets along basement/sediment and granite/sediment contacts at meso- to epizonal levels, forming cupolas in positions controlled by

D3-structures. Most plausibly, their origin may be modelled as being due to orogenically-induced crustal thickening and re-melting of residues of earlier granite extraction. If this alone explains their geochemical speciation, or if fractional crystallization at depth must be invoked (Lehmann and Lavreau, 1988), remains an unsolved puzzle as for tin granites worldwide. It is interesting to note that in several, though not all, aspects they are comparable to the Hercynian tin granites of Europe and of northwestern Africa.

An important feature of the Kibaran granitoids in the area is their frequent occurrence in well-defined composite plutons. These may comprise cores of pre-Kibaran basement as well as Kibaran high-grade rocks, and they then resemble mantled gneiss domes. Anticlinorial or tectonically elevated situations are the rule, and many of these have been sites of intrusive activity from G1 to G4 (except, of course, the rift-related G3-granites). As earlier granites are roughly concordant with the country rocks, preferential erosion of the intrusive rocks led to the formation of wide circular depressions rimmed by more resistant metasediments; these are the spectacular “arenas” so typical for the Kibaran in Tanzania and in Uganda, but absent in the morphologically immature Congo–Nile water-divide of Burundi and of Rwanda.

With the Lomamian Orogeny (D3) marked by tin granites, and the associated typical “Kibaran mineralization”, the evolution of the Kibara belt ends. All later events, among them the intrusion of alkaline complexes and of carbonatites (750–700 Ma, 500 Ma, etc.), the Pan-African resetting of mineral ages (600–500 Ma), and the formation of the Central African Rift (ca. 30–0 Ma), are not restricted to the Kibara belt, but concern much wider areas of central and southern Africa.

2.4. Geodynamic models

No field evidence is available to suggest the involvement of oceanic crust in the evolution of the Kibara belt. There are no ophiolites, neither orogenic andesites nor diorites (including other calc-alkalic magmatic rocks of the typical sub-

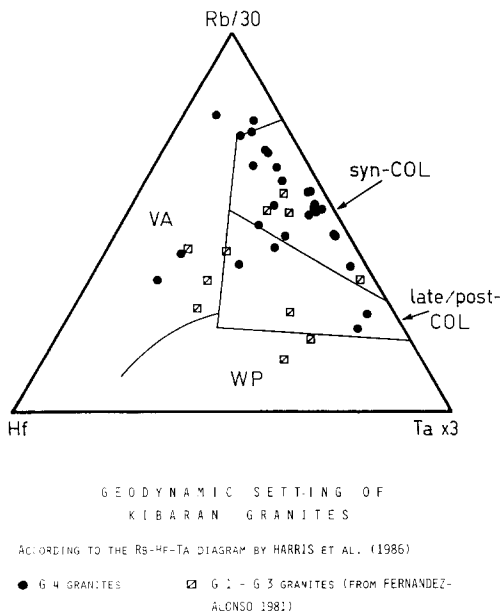


Fig. 6. Geodynamic setting of Kibaran granites (Guenther et al., 1989). Based on the discrimination diagram by Harris et al., 1986. VA—volcanic arc granites; WP—within plate granites; syn-COL—syn-collisional granites; filled circles are G4-granites; open squares are G1- and G2-granites.

duction suite), and there is no indication of paired metamorphic belts. In contrast to these findings, some amphibolites in the granite-gneiss domes have geochemically an oceanic affinity (Jung and Meyer, 1990). Most probably, these magmas are related to early Kibaran rifting or crustal thinning, stopping well before the formation of an ocean. As also described in the introduction, there is a very strong case for a wholly intra-continental evolution of the Kibara belt.

Compared with younger orogenic belts, like the European Variscides or the Alps, the time covered by Kibaran evolution is extremely long — about 500 million years — and our ability to detail different stages of this evolution is rather poor. This may have been one of the reasons for Cahen et al. (1984) to limit the term “Kibaran” to the time span from about 1400 to 1100 Ma, and to attribute the further history of the area to the Pan-African cycle. However, this overlooks the individual nature of the belt continuing until about 900 Ma, as the typical “Kibaran” tin granites with their important mineralization are spatially restricted to the Kibara belt, and are an integral part of its evolution.

Clearly, there is an overlap in time between Kibaran and Pan-African evolution. The reason for this is the protracted history of crustal convergence at the southern and eastern margins of the African continent, from about 1300 to 500 Ma.

A further interesting aspect of this is the repeated extraction of granitic melts, supposedly from the same source rocks at depth (Archean and/or Lower Proterozoic) without the addition of new supercrustal material (there is no evidence of large-scale subduction of Kibaran sediments). In this respect, the Kibara belt could, in future, be an interesting region for research on restites and restite-derived melts.

The inferred intracratonic setting of the Kibara belt raises the question as to whether it may be considered as an aulacogen related to Wilson cycle orogenic belts in southern Africa and adjoining parts of Ur-Gondwana (Hartnady, 1991). Kibaran mineralization is atypical of aulacogens (compare Erikson and Chuck, 1985), however, and present knowledge of the overall

structural setting of the belt is too restricted to allow a rational discussion of this point.

3. Mineralization

In the past, mineral deposits of Sn, W, Nb/Ta and Au in the area have been discovered most efficiently by panning alluvial material and by a physical search for mineralized outcrops (U, Be, Li, W). More recently, geochemical and geophysical methods have increasingly been used, and this has resulted in the recognition of the mafic/ultramafic belt in Burundi with its Ni, Co, Cu, Pt, Cr, Ti, V and Fe mineralization (Radulescu, 1982) and equivalent prospects in Tanzania.

The currently known mineral inventory (Ziserman et al., 1983; Barnes, 1961; Tissot et al., 1982; Pohl, 1987a; Pohl, 1992) allows the following generalized synthesis of geological evolution and mineralization.

3.1. Mineralization syngenetic with the Kibaran sediments

There are no records of any production of syngenetic minerals from the Kibaran. Earlier suggestions that some gold placers may have been derived from low-grade primary sources in Kibaran meta-sediments or volcanics could not be confirmed; all primary sources were recognized as epigenetic (see for example Pohl and Hadoto, 1990).

Lenticular stratiform bodies of magnetite (\pm pyrite) and specular hematite (for example Kamena/Uganda within the Middle Group) which are currently uneconomic because of the small quantities involved occur in quartzites and metapelites. Such stratiform bodies are being increasingly recognized as magnetic surveys are carried out, although their origin has not yet been investigated. It is tempting, however, to consider these ores as syngenetic hydrothermal exhalites, distally related to volcanism. This may indicate a potential for economically more interesting base-metal mineralization.

A further encouragement to search for synse-

dimentary hydrothermal ores is the discovery of manganiferous garnetites in western Rwanda (Pohl, 1987b), as extensive Mn halos are characteristic features of this type of deposit.

3.2. Mineralization associated with the 1275 Ma mafic layered intrusions

Geochemical surveys prompted by earlier reports of Ni mineralization in southeastern Burundi led to the discovery of Ni-laterites of possible economic size and grade at Musongati (Radulescu, 1982). Meanwhile, many other mafic/ultramafic bodies have been located within a belt up to 50 km wide and nearly 400 km in length extending from the eastern shore of Lake Tanganyika through Burundi to the Tanzanian/Ugandan border.

Anorthosites and leucogabbros of the complexes may contain lenses of vanadium-bearing magnetite, Ti-magnetite and ilmenite (Buhoro/Burundi). Sulphide mineralization with Ni, Co, Cu, platinum group metals and some chromite has been found in ultramafic members (pyroxenites, periodites, dunites) and their immediate country rocks in Burundi (Musongati, etc.: Niyonzendo, 1984; Jedwab, 1987) and in Tanzania (Kabanga: Van Straaten, 1984; Gosse, 1992). Magmatic stratigraphy and mineralization are quite similar to higher parts of the Bushveld igneous complex (Deblond, 1990).

The Ni(-Cu-Co)-sulphide deposit at Kabanga is, at present, the most interesting potential mine in the Kibara belt. From a magnetic anomaly known over 8 km length, only about 1000 m have been drilled until now. Reserves thus established comprise 11.7 Mt of ore at 1.72% Ni, 0.26% Cu and 0.12% Co, at a cutoff grade of 1% Ni. The Kabanga ultramafic body is intrusive into low-grade Kibaran clastic metasediments, on the eastern flank of an overturned anticline (Fig. 7). It has the shape of a keel and consists broadly of a peridotitic core surrounded by pyroxenite and an outer gabbro shell. Adjoining phyllites suffered hornfels-metamorphism. The ultramafic rocks consist of granular black serpentinized olivine, light-green ovoid poikilitic orthopyroxene, and brown-green or-

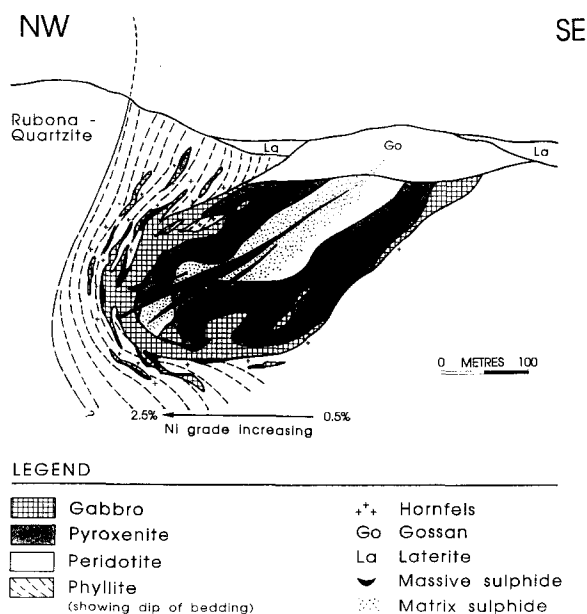


Fig. 7. Section showing main features of the Kabanga Ni-Cu-Co deposit, Western Tanzania (after Gosse, 1992).

thopyroxene lathes. Clinopyroxene is rare, plagioclase and pyrrhotite are ubiquitous as xenomorphic intra-cumulus phases.

The mineralization comprises massive sulphide dykes and layers, mainly in the ultramafic core (Fig. 7), but also in the gabbro and in the metasediments. Disseminated sulphides occur only in the ultramafic rocks, where they form a three-dimensional interconnected sulphide matrix when the sulphide content exceeds 25%. The predominant sulphide mineral is pyrrhotite, with minor amounts of pentlandite, chalcopyrite, cobaltite and pyrite. Magnetite and chromite are accessory phases, while gold and PGM contents are below 0.35 ppm (Gosse, 1992).

In the absence of more detailed data, the origin of the deposit may tentatively be described in terms of the interplay of unmixing and segregation of silicate and sulphide liquids within the crystallization sequence of the intrusion. More rapid cooling of its margins apparently produced the gabbro cusp first, which then collected heavier cumulus phases, and sulphide liquids formed in higher parts of the intrusion that are now eroded away.

3.3. Mineralization associated with the tin granites

Essentially, it is this group that is the base of historic and present mining within the Kibara belt. Accordingly, a large number of publications concerned with different aspects of the deposits is available.

Four subgroups of G4 granite-associated mineralization can be differentiated:

—Pegmatites with Sn and Nb/Ta (Li, Be, U/Th) and the nonmetallics muscovite, quartz, feldspar, kaolinite;

—Quartz veins with Sn and W (pyrite, siderite, Bi, Au, U);

—Talc deposits developed by hydrothermal alteration of dolomites and/or mafic magmatic rocks; and

—Auriferous silicification zones, breccias and quartz veins with native Au, pyrite and arsenopyrite, or with magnetite, specularite, and in adjoining Kivu with PGM.

The **Kibaran pegmatites** have been studied by numerous authors from mineralogical, geological and metallogenic points of view. Some of the pegmatites have an exceptional number of rare minerals, especially those cutting basic country rocks (for example, about 100 have been described from Buranga/Rwanda: Bertossa, 1965; Von Knorring, 1969). With the exception of mineralogical observations, however, the pegmatites have not been investigated in detail since Varlamoff (1972, 1975, and many earlier publications).

Varlamoff (1975) observed a pronounced zonation of different types of pegmatites and associated quartz veins in relation to granite cupolas. The large pegmatite fields, however, ordinarily lie apart from big vein-type deposits, which is probably due to different intrusion levels (meso- versus epizonal) and consequently different de-volatilization processes of the respective source granites. The mineralized pegmatites are closely related to the LCT-type (characterized by elevated lithium-caesium-tantalum) after Cerny (1991). Many of the pegmatites are strongly argillized. This may be an effect of supergene alteration in some cases. However, un-

altered pegmatite bodies close to completely kaolinized ones, make a hydrothermal alteration the more probable alternative. Argillization facilitates exploitation of low-grade orebodies, where cassiterite and columbo-tantalite occur disseminated in the rock. High-grade ore shoots are normally close to quartz cores or to contacts.

The **Sn- and W-quartz veins** are the major source of cassiterite in the area, and of practically all of the tungsten ore. Important deposits are normally made up of a multitude of single veins and veinlets forming stockworks and vein fields, whose position is controlled by anticlines or domes which apparently acted both as preferred intrusion sites of G4-granites and as anticlinal fluid traps. As a rule, deposits are hosted by Kibaran metasediments; exploitable occurrences within granites are extremely rare. The major vein fields are clearly separated from pegmatite districts; some pegmatites, however, have been observed to be cross-cut by mineralized quartz veinlets, and others, as at Musha/Rwanda, pass upwards into a zone of chaotic quartz-muscovite-kaolinite veins and pockets with cassiterite (Varlamoff, 1969). Major tin fields are clearly separate from tungsten deposits, but there are numerous quartz veins with both cassiterite and wolframite (Baudin et al., 1984), and small amounts of cassiterite have been found in ferberite deposits (Pohl, 1975). Both cassiterite and wolframite veins are known in host rocks ranging from quartzite to metapelites, but in the largest tin district of the area, Rutongo/Rwanda, quartzites are preferentially mineralized; in the same region, carbonaceous psammites and schists, for example at Shyorongi, are characteristic hosts for tungsten mineralization (Fig. 8).

The spatial separation and differing host lithology of important tin and tungsten districts have been used as arguments for syngenetic models of the tungsten ores, in contrast to the generally accepted epigenetic derivation of cassiterite from G4-granites. To allow a more detailed discussion, typical features of both will be summarized below.

The tin district of **Rutongo** (Pohl, 1978) occupies the eastern limb of a large anticlinorium, which plunges at a low angle to the NNW. The

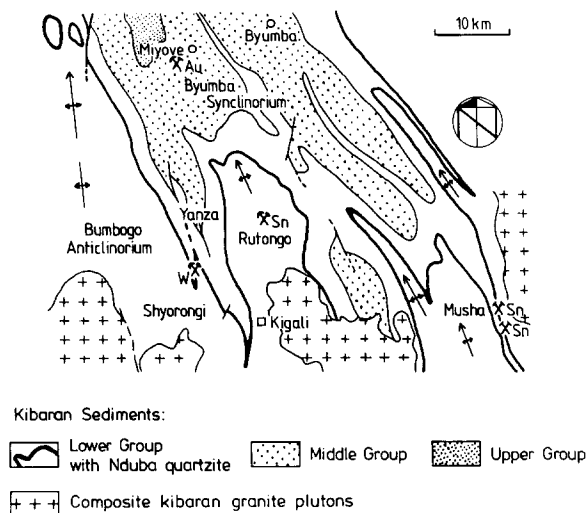


Fig. 8. Geological setting of major tin, tungsten and gold deposits in central Rwanda (Pohl, 1977).

western flank is steeply dipping and locally overturned, while the eastern limb dips about 30° to the east. The folded rocks are a suite of metapelites, metasandstones and quartzites of the Lower Group. The metamorphic grade increases to the south, where the core and eastern limb of the anticline are intruded by the Kigali G4-granite exposed in a morphological depression to the southeast (Fig. 8). There, outcrops show white sericitized and strongly kaolinized, tourmaliniferous, medium-grained granite. This is due to hydrothermal alteration (Pohl and Guenther, 1991).

The cassiterite-quartz veins at Rutongo occur almost exclusively in thicker bands of competent quartzites and metasandstones interbedded with metapelites. The latter did not fracture, and thus contain very few veins. Single veins have an average thickness of 1 m, ranging from several cm to more than 10 m. They form fields of parallel veins comprising up to several hundred single veins. Productive zones are those where the combined thickness of the veins is larger than 3.5–5 m measured in 50 m sections at a right angle to the strike of the veins. Five productive zones each consisting of several vein fields can be differentiated in the Rutongo district; the most important one lies just to the east of the major

N–S-trending overthrusts near the crest of the anticline. More than 80% of the district's cumulative cassiterite production of about 30,000 m.t. originated from this zone, which is up to 1500 m wide and over 6 km long in a N–S direction. The veins also strike N–S and dip steeply to the west.

Whereas the large Rutongo anticline is essentially a product of D2-folding along NNW–SSE-trending axes, a later compressive event (D3-Lomamian) is responsible for a further tightening of the anticline producing the overthrusts of N–S strike and easterly dip, and the tensional fractures controlling mineralization (Pohl, 1977). The syntectonic nature of the veins and their strict symmetry relationship with the anticline had been recognized earlier (Aderca, 1957) without, however, differentiating the two deformation phases.

The veins contain quartz, muscovite and cassiterite; in addition, minor amounts of kaolinite (partly after feldspars), tourmaline, rutile/ilmeno-rutile, arsenopyrite, pyrite, chalcopyrite, galena and traces of gold (in cassiterite placers) have been found. Although most veins consist of massive “buck” quartz indicating simple vein growth, some allow differentiation between brecciated early white quartz which is cemented by grey quartz with more sulfides. Hematite/goethite coatings along late joints cutting the vein quartz are ubiquitous. Country rock alteration includes sericitization, tourmalinization, kaolinization and silification; disseminated cassiterite is rare. Most cassiterite occurs with coarsely crystalline muscovite at vein contacts, while the milky white or grey vein quartz is often nearly barren.

Fluid inclusion studies (Pohl and Guenther, 1991) show that early vein-forming fluids (Sn1 of Fig. 9) at Rutongo had moderate salinity and high gas contents (CO_2 , CH_4 and N_2); homogenization temperatures (T_h) cluster between 250° and 380°C . These fluids are best understood as the product of the devolatilizing G4-granite at depth. The fluids boiled when entering the opening spaces of the deposit at an estimated depth of 2000 m, thus inducing deposition of cassiterite. The following fluid generation (Sn2) had slightly lower T_h , but higher salinities, probably due to

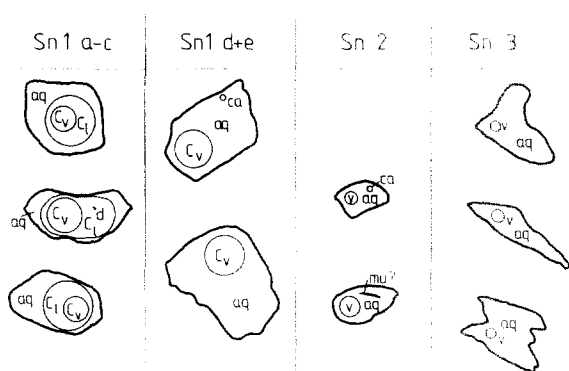


Fig. 9. Fluid inclusion types in quartz and cassiterite from Rutongo/Rwanda (not to scale; inclusion sizes vary from <5 to 100 μm). *aq*—aqueous fluid; *Cl*—liquid CO_2 ; *Cv*—gaseous CO_2 ; *v*—vapour bubble; *ca*—carbonate daughter mineral; *mu*—muscovite; *d*—undetermined daughter mineral.

hydrothermal alteration of solidified granite at depth. Further cooling of the granite by meteoric waters produced aqueous fluids of continuously decreasing temperatures (Sn3).

Three important tungsten mines in Rwanda (Nyakabingo–Shyorongi, Gifurwe and Bugarama) and Kirwa in southwest Uganda are aligned along the NNW-trending Bumbogo anticlinorium (Fig. 8). Within this large structure, second-order anticlines controlled the location of mineralization. The roughly linear arrangement, in conjunction with a similar stratigraphic position within the upper part of the Lower Group, has prompted the designation “tungsten belt” (De Magnée and Aderca, 1960).

The orebodies at **Shyorongi** (also called **Nyakabingo**) lie within a domal structure formed by the rectangular intersection of D3-crossfolds with the main D2–D3 fold axes similar to Rutongo. The orebodies consist of concentrations of quartz veins, which demonstrate structural control by a compressive stress regime (Fig. 10). Host rocks are pyritiferous black shales and turbiditic grey psammities of the upper part of the Lower Group, immediately below the conspicuous Nduba quartzite marker bed. A few km to the south of the mine, strongly kaolinized G4-granite can be observed in outcrops (Fig. 8).

Mineralization consists mainly of porous, vuggy reinite (ferberite after scheelite) in quartz(-muscovite) veins. These occupy both

bedding plane joints in the turbidites as well as cross-cutting fissures. As a rule, the first are cut by the latter, indicating early high fluid pressures which were then released by fracturing; the reverse relationship, however, can also be observed. Thickness of the quartz veins is mostly 10–30 cm, in places reaching 1 m and more. They contain earlier milky and later grey quartz. Some very thin veins in black shales consist of fibrous quartz. Apart from reinite, platy ferberite, quartz and muscovite, minor amounts of scheelite, tungstite, anthoinite, cassiterite, kaolinite, minute needles of tourmaline, arsenopyrite and pyrite have been found. Late impregnations and coatings of goethite or hematite are ubiquitous. Country rock alteration is macroscopically not visible, but includes tourmalinization, silification and sericitization. Within the mineralized zone, pyrite is almost completely leached from the black shales, leaving a rock studded with nearly empty cavities. The vertical spread of intense quartz veining and mineralization is restricted to about 150 m. The mine's cumulative production (1952–1985) amounts to about 3500 m.t. of concentrate, remaining reserves may be equal to that figure.

An interesting feature of several Kibaran black shale-hosted tungsten deposits are beds studded with cm-sized tungsteniferous nodules, which were called “**ferberite nodules**”. They occur at Shyorongi and at Gifurwe (Frisch, 1975) in Rwanda and at Nyamulilo and Ruhizha (Pargeter, 1956) in Uganda. At Gifurwe, these flattened nodules consist of clay minerals, sericite, quartz, graphitic matter and iron-hydroxides (at least in part after pyrite); they contain about 1100 ppm W (11 samples). Apparently similar nodules at Ruhizha contain more ferberite and were actually exploited as low-grade ore (Pargeter, 1956). Microveinlets of quartz and tourmaline were occasionally observed. The nodules form horizons in black shales and are clearly stratabound. Most probably, they only contain tungsten in areas of intense quartz veining, but this remains unconfirmed, as comparable nodules outside of the mines have not been found until now, due to poor outcrop conditions. At Ruhizha and at Nyamulilo the nodules were ear-

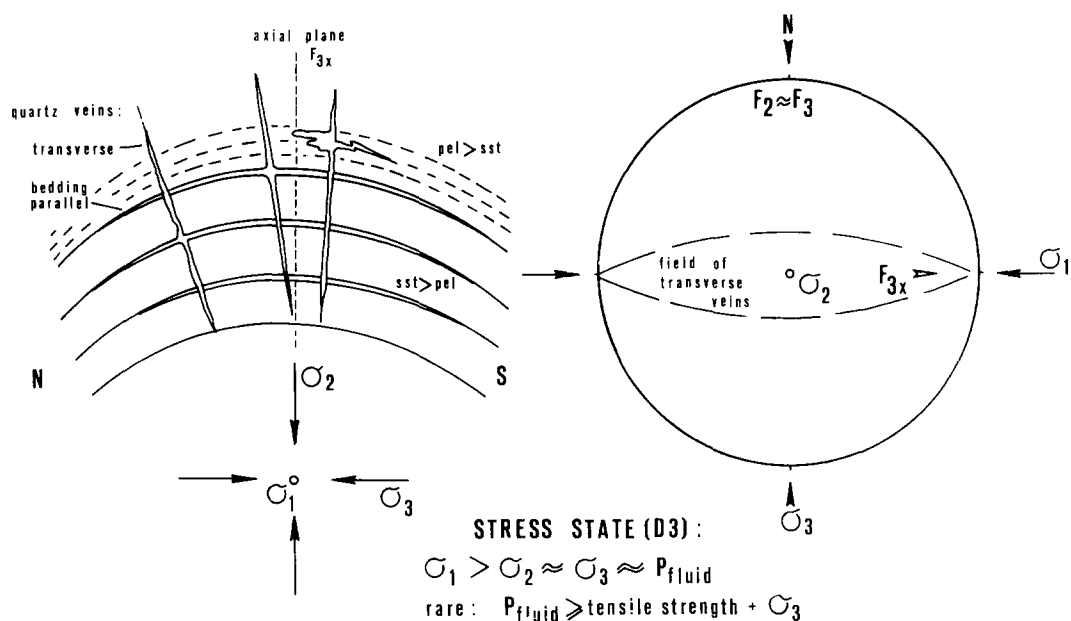


Fig. 10. Schematic section and stereographic projection of structures and tungsteniferous quartz veins in Rulizahene mine, Nyakabingo, Rwanda, as a base for stress field reconstruction. *pel*—black shales; *sst*—meta-psammities.

lier thought to have trapped tungsten diagenetically, which then would have been partially mobilized by metamorphism and/or granite derived fluids to produce the main mass of clearly epigenetic ores (De Magnée and Aderca, 1960). Now, they are interpreted as diagenetic concretions (originally phosphatic?) that were preferential sites of W precipitation from pervading hydrothermal solutions (Frisch, 1975).

Fluid inclusion research (Pohl and Guenther, 1991) reveals that the evolution of tungsten-depositing fluids was nearly identical to those at Rutongo. Fluid type W1 comprises mixtures of an aqueous phase of moderate salinity with a $\text{CO}_2(-\text{N}_2-\text{CH}_4)$ -phase that is either predominantly liquid or gaseous at room temperature (Fig. 11). W1a and W1b homogenize into vapour and W1c–e into liquid state at about 300°C . The syngenetic trapping of high- and low-density variants of W1-fluids indicates unmixing (“boiling”) as the fluids expanded into low-pressure spaces at the site of mineralization. Physical trapping conditions may be estimated at $300^\circ\text{--}500^\circ\text{C}$ and 2000 to 400 bars. W2-inclusions are aqueous and more saline, and contain calcite and

muscovite daughter crystals. W3-fluids are aqueous and have low salinities. While W1-fluids may have exsolved from the crystallizing G4-magma, it is thought that W2 and W3 represent fluids which have interacted with already-solidified granite at depth. In addition, (trace) lead isotope ratios in sulfides from Rutongo and from Nyakabingo are comparable with each other and with pyrite from Kibaran black shales far from mineralization (Fig. 12), thus confirming the close genetic relation between tin and tungsten ores as well as the interaction of granite-derived fluids with the country rocks.

At Kibanda/Kibuye/Rwanda a talc deposit was recognized during the UNESCO Geotraverse in 1986 (Pohl, 1986). Scientific investigations and drilling soon followed (Prochaska and Rulinda, 1989). The deposit lies in a suite of metasediments and metadolerites of the Middle Group of greenschist-facies metamorphic grade. At Kibanda, a 50-m-thick lens of dolomitic carbonate is sandwiched between two metadolerite members. The carbonate was transformed by hydrothermal solutions into a talc (-carbonate) body separated from the metadolerite by a shell

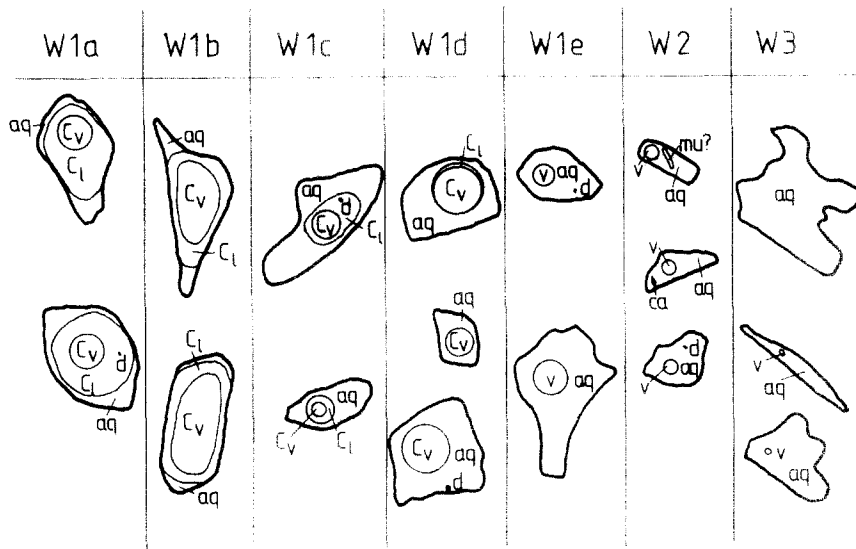


Fig. 11. Fluid inclusion types in quartz from the ferberite deposit at Shyorongi/Rwanda (not to scale; inclusion sizes vary from < 5 to 100 μm). For abbreviations see Fig. 12.

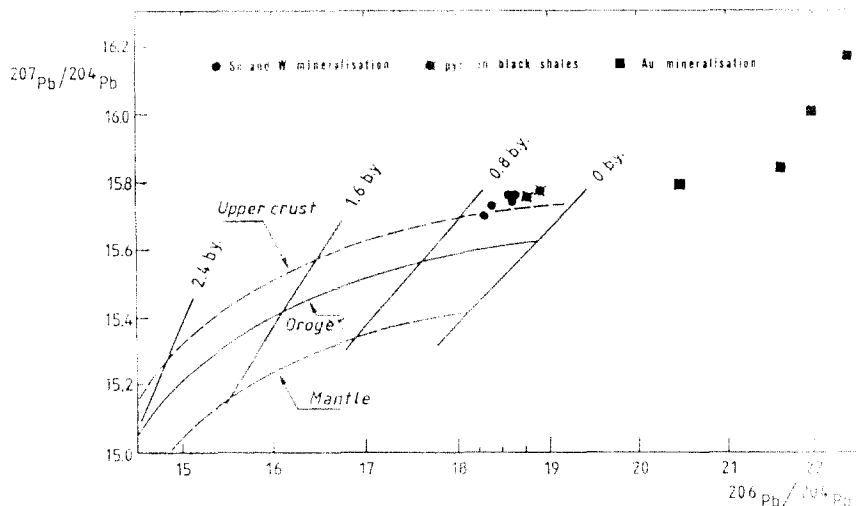


Fig. 12. Trace lead isotope data from pyrite samples of Kibaran mineralization (Cauet and Pohl, 1988). Evolution curves and isochrons are those generated by the plumbotectonics model of Doe and Zartman (1979).

of chlorite schists. The alteration seems to be related to minor quartz-muscovite veinlets; fluid inclusions in the quartz are highly saline and homogenize at about 240°C. The formation of talc comprised essentially only introduction of silica into the dolomite and dispersion of Ca. It is highly probable that this process is related to

movement of fluids associated with G4-granites known in the area.

Auriferous quartz veins and breccias with secondary iron hydroxides in the superficial alteration zone and primary pyrite and arsenopyrite (locally also with magnetite, specularite) at depth are the source of Au placers, which have

contributed practically all the gold exploited in the area. Many of the primary sources have been located (Ziserman et al., 1983; Niyondezo, 1984; Barnes, 1961), but most of these were found to be of uneconomic low grade and small size. Only locally, high-grade ore shoots are exploited by small-scale mining, the largest mine at present being based on the vein at Baradega/Miyove in Rwanda (Fig. 8).

Auriferous quartz veins occur typically in clastic Kibaran metasediments, while the iron-rich gold-bearing breccias are restricted to basic metavolcanic country rocks. The first are characterized by a gangue of quartz, sulfides, tourmaline, rutile and muscovite, while the second differ mainly by a predominance of Fe-minerals, including sulfides, magnetite and specularite.

Important gold districts are Nyungwe (Rwanda) and Mabayi (northwest Burundi), Miyove in Rwanda, and a wide area extending from there northwards to the outlier of Kibaran sediments (quartzites, sandstones and pelites) in the Buhweju plateau (Fig. 2). At Buhweju, the placers nearly always contain some cassiterite and monazite. In this area, small auriferous quartz veins are not infrequent in both the underlying Buganda–Toro schists and in the Ki-

baran suite; pyrite, and more rarely Cu, Pb and Zn sulfides occur in the veins (Barnes, 1961). At the southern Buhweju escarpment, near the gold placers of Mashonga, leucocratic muscovite granite (G4) intrudes Buganda–Toro rocks; its alteration halo with tourmalinization and quartz veining affects basement gneiss and schists, but also the basal Kibaran conglomerate (Fig. 3). Farther south in the southwestern Kigezi district, some gold has been found in limonitic (originally sideritic) veins with bismutite, pyrite, wolframite and cassiterite (Pargeter, 1952), a type of mineralization which is similar to the NW-Burundian breccia zones (compare Brinkmann et al., 1994, this issue).

The regional distribution of the gold districts (Fig. 2) is generally away from granite domes and peripheral to the Sn–W zones, although transitional overlap — mainly with ferberite — clearly occurs. This was earlier interpreted as indicating the gold zones to be a more distal and lower-temperature “epithermal” phase of mineralization in relation to the tin and tungsten ores (Frisch, 1975). Meanwhile, fluid inclusion studies have shown that most gold quartz veins, as well as the breccias, formed at even higher temperatures (about 450°C) than tin and tungsten,

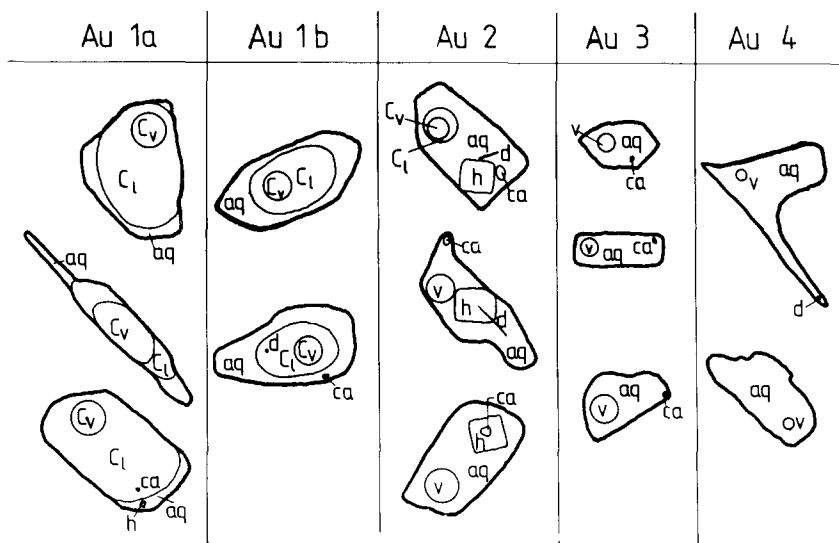


Fig. 13. Fluid inclusion types in quartz from gold deposits in Rwanda and Burundi (not to scale; inclusion sizes vary from <5 to 100 μm). For abbreviations see Fig. 12.

and at pressures of about 2 kbars from fluids with about 8% NaCl equivalent and high contents of liquid CO₂ (inclusion types Au1a and Au1b of Fig. 13) which unmixed (“boiled”) when entering depressurized sites of deposition. The highly saline Au₂-inclusion types (Fig. 13) are thought to represent the solute-enriched aqueous end-members of this unmixing process. Major solutes comprise Na–K–Ca (–Fe–Mn)–Cl, probably strongly dependent from interaction with country rocks. Au₃- and Au₄-inclusions are aqueous and have decreasing salinity and homogenization temperatures. The CO₂-phase of the Au-fluids does not contain N₂ and CH₄ setting them apart from Sn- and W-fluids. In addition, trace lead in sulfides from these deposits is highly radiogenic J-type (Fig. 12). A direct derivation of fluids depositing distal Kibaran gold from G4-granites is, therefore, rather unlikely. An exception are gold deposits closely associated with G4-granites as at Mashonga (Fig. 3), where the fluids are indistinguishable from those forming the tin and tungsten ores.

Accordingly, Kibaran gold ores and low-grade sources of placers in the area encompass a variety of types, including G4-granite-derived ores (\pm ferberite, cassiterite), and others apparently unrelated to granites. Tectonic control of gold deposits (Franceschi, 1990), however, is so similar to tin and tungsten, that a roughly contemporaneous formation is strongly indicated; an observation confirmed by isotope dating (Brinkmann et al., 1994, this issue). This imposes constraints on the genetic interpretation which will be discussed later.

3.4. Mineralization related to alkaline complexes and carbonatites

These are post-Kibaran deposits related to magmas of sub-crustal derivation, typically present in consolidated cratons.

At **Lueshe** in eastern Zaire, carbonatites and alkali-syenites intruded at around 500 Ma contain pyrochlore that is exploited for its Nb content from the lateritic cover (Maravic et al., 1989).

The small bastnaesite deposit at **Karonge** and

minor equivalents elsewhere in Burundi are probably also associated with the Pan-African intrusive activity (Niyondezo, 1984). At **Karonge**, bastnaesite in veins and veinlets forms orebodies of a stockwork type. It is accompanied by monazite and a gangue comprising quartz, barite and goethite. A hydrothermal derivation from a hidden alkali-carbonatite plug is assumed (Van Wambeke, 1977).

It should be stressed that the intermittent intrusion of magmas related to crustal extension during the Pan-African period in the area, from about 700–500 Ma, indicates a cratonic and not an orogenic setting. Accordingly, we believe that the definition of a “Katangan” Western Rift mobile belt (Cahen et al., 1984) is erroneous.

5. Discussion and conclusions

An intracontinental setting for the Kibara belt throughout its evolution (Klerkx et al., 1984), due to interaction between the Congo (Zaire) and Tanzania cratons, appears quite well documented. It was preceded by Eburnean orogenic activity and followed by repeated movements and mantle activity of anorogenic type during the Pan-African, which is now thought to comprise the period from about 950 to 450 Ma (Kröner, 1984). Opposing views have been published, however, concerning the attribution of the G4-granites (1000–950 Ma) and the roughly time-equivalent Lomamian folding to either a late Kibaran (Pohl, 1977) or an early Pan-African phase (Cahen et al., 1984). Generally, the folding of molasse deposits is not questioned as representing a final stage of orogeny. The large time lag between the main orogeny (about 1300 Ma) and this last compressive phase in the Kibaran is certainly unusual. We suggest that this is related to the easterly drift of the orogenic climax throughout Kibaran and Pan-African time in eastern Africa (Kibara belt–Irumides–Lurio belt–Mozambique belt). The Lomamian compressional phase in the Kibara belt must be understood as a “far field effect” of Mozambiquian collision farther east, between an African and an Indo-Antarctic craton.

In view of their metallogenic significance, the origin of the G4 “tin” granites needs clarification. These specialized granites plot as volcanic arc and syn-collisional granites in tectonic discrimination diagrams, and are similar to many post-orogenic granites (Guenther et al., 1989). They are not A-type and a crustal derivation cannot be doubted for those east of the rift. All this, and the rarity of contemporaneous basic magmatic activity, appears to favour an origin by stacking of continental crust similar to the late-Hercynian “tin” granites in central and western Europe (Windley, 1984). It is proposed that the geological event responsible for renewed widespread crustal melting throughout the Kibara belt is compressive deformation and resulting crustal thickening caused by the Lomamian folding phase defined in the Kibaran foreland. Apart from deforming Kibaran internal and external molasse, this phase partly tightened or steepened earlier structures within the belt. Elsewhere, new folds (F3) developed at an angle to older ones (F2). The resulting tectonic highs at anticlinal intersections were preferential sites for tin granite cupolas. At appropriately oriented earlier faults and basement structures, shearing was induced.

Mineralization associated with the tensional (rift and geosynclinal) stage of the Kibara belt is restricted to minor iron ores and manganese-rich beds. Prospecting until now has largely ignored this environment, but the increasing recognition of volcanic rocks in the sedimentary pile should encourage a search for stratabound ores.

The mafic-ultramafic belt of Burundi and Tanzania, with its linear structural control and its layered mafic intrusions, including the typical mineralization (Ni, Cu, Co, Fe, V, Ti), represents a major rifting event following the main orogenic deformation of the Kibaran rocks.

The origin of the tin–rare-metal pegmatites and of the majority of metalliferous quartz veins is clearly associated with the G4-granites. A strong structural control of both intrusion sites of the granites and the mineralization related to a compressive deformation phase has generally been documented. In some respects, earlier metallogenic models (Varlamoff, 1975; Frisch, 1971,

1975; Pohl, 1987a) appear too schematic, however, and have to be revised in view of the more recent data.

The fluids involved in Kibaran Sn–W–Au mineralization have meanwhile been investigated (Pohl and Guenther, 1991). In addition, the mineralogical composition of the ores and of associated hydrothermal alteration can be used to characterize the metallogenic environment. Most importantly, fluorine minerals are quite rare in all types of mineralization. The greisens described in earlier publications on the Kibaran consist of muscovite-quartz aggregates. Topaz and fluorite are so infrequent (Bertossa, 1968) that the fluids would, in most cases, have contained little fluorine. Apatite and other phosphates are much more frequent (Bertossa, 1968), which suggests that P was one of the more important volatiles, together with H₂O, boron (tourmaline is ubiquitous) and, of course, Cl and CO₂. High Na-contents of early magmatic fluids are indicated by frequent albitization of G4-granites and pegmatites; expelled ore-forming fluids trapped in Sn and W veins have only moderate NaCl-contents. In line with the present understanding of fluids involved in this type of mineralization (Campbell et al., 1984; Manning, 1984; Roedder, 1984), a considerable variability in time has been found in single deposits. This may be interpreted as reflecting early segregation of magmatic fluids, followed by later fluids of external derivation that have interacted with granites and country rocks at different fluid/rock ratios and at falling temperatures.

The question as to the original source of the metals remains a matter of speculation in the Kibara belt as elsewhere. Because of the relatively high tungsten contents of Kibaran metasediments in southwest Uganda (Pargeter, 1956; Jeffery, 1959), a synsedimentary deposition of tungsten has been proposed by some authors to have preceded concentration by metamorphism and/or granite intrusions (Pargeter, 1956; Jeffery, 1959; De Magnée and Aderca, 1960; Reedman, 1967). In view of the very low grade of metamorphism and the absence of quartz veins in unmineralized areas, however, a lateral secretion model is unlikely. Furthermore, there is no

indication of "granitization" in the immediate surroundings of the ferberite deposits. Accordingly, the carbonaceous shales with high tungsten contents and ferberite nodules which are characteristic country rocks of ferberite quartz vein fields cannot be the source of the tungsten in the deposits. On the contrary, these anomalous compositions represent geochemical haloes of epigenetic mineralization, as shown by Frisch (1975) at Gifurwe.

Anomalous geochemical trace contents of tin in G4-granites (Lowenstein, 1966; Lehmann and Lavreau, 1988) and the geological environment suggest that the granites (and not country rocks leached by hydrothermal convection systems) were the immediate source for the ores. Accordingly, those crustal rocks that were affected by partial melting to produce the tin granites were most probably the original source. The previous geological history of this deep basement would then have determined the availability of the metals, and at least of part of the volatiles. Whether this involves an earlier geochemical enrichment, as demanded by many metallogenic models (Routhier, 1980), or rather more effective liberation of the metals from rocks with a normal crustal concentration (i.e., restite re-melting, etc.), or only prolonged magmatic differentiation of granitoids (Lehmann and Lavreau, 1988)

remains one of the fundamental problems of metallogeny.

Many Kibaran gold deposits have a geological setting clearly different from tin and tungsten veins; they appear in synclinoria rather than in anticlinal positions, their structural control is more typically brittle faulting and shearing than folding, and they attain higher lithostratigraphic levels than tin and tungsten which are restricted to lower parts of the Kibaran sedimentary pile. Most importantly, many occur at considerable distances from G4-granites and, consequently, their origin can hardly be sought in volatiles and fluids segregated from these magmas. Furthermore, fluid inclusion studies reveal that these gold quartz veins and breccias were formed at higher temperatures and pressures compared with granite-derived tin and tungsten.

A model reconciling the contradictory findings that gold may both be associated with tin granites and their rare-metal deposits, as well as occur far from these intrusives, has to consider the processes taking place in the crust underneath the Kibaran orogen thickened by Lomaginian deformation. There, thermal equilibration of the crust and radiogenic heating must have combined to create a prograde metamorphic front progressing upwards (Fig. 14). Anatexis was conducive to melting in the lower crust, thus producing the magmas whose most evolved

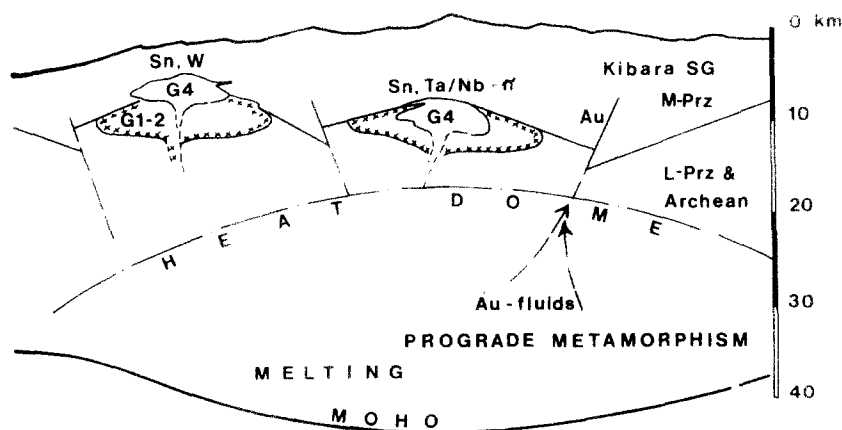


Fig. 14. Conceptual model of origin of Kibaran tin granites and gold mineralization: both G4-granites and gold depositing fluids are generated at depth as a final consequence of crustal thickening.

members are the tin granites. At the same time, devolatilization of lower crustal rocks produced fluids that were channelled into permeable structures along deep faults separating basement blocks. Gold and quartz were deposited where these fluids started to boil.

Accordingly, both the anatectic melts and the gold ore-forming fluids are thought to have been generated at depth as a final consequence of crustal shortening and thickening caused by the Lomamian Orogeny. The source of the gold may be sought either in deeply-buried Archean greenstone belts or in Lower Proterozoic mafic rocks underneath the Kibaran metasedimentary pile, both containing many gold deposits east and south of the belt. This deduction is supported by the co-occurrence of alluvial gold and platinum (plus cassiterite) in Kivu/Zaire (Jedwab, 1992), as elevated PGM-contents strongly indicate ultimate source rocks of mafic or ultramafic composition.

5.1. Some suggestions for exploration

Some exploration approaches derived from the present metallogenic analysis of the Kibaran in Central Africa are:

—A search for industrial minerals and rocks (already known to occur are andalusite, feldspar, muscovite, kaolin, quartz, talc, wollastonite; playa sediments of internal molasse basins may contain other minerals).

—Undoubtedly, more tin and tungsten deposits can be found. Possible ore types are proximal skarn ores in contact metamorphic carbonates, disseminated Sn–Nb–Ta within, but near the roof of granite cupolas, and scheelite in mafic rocks (distal skarn ores).

—Sedimentary exhalative or volcanogenic hydrothermal base-metal ores should be searched for where synsedimentary volcanism, tensional tectonics and rapid basin subsidence coincide. Basin analysis based on the refined lithostratigraphic classifications should be applied for this purpose.

Acknowledgements

The scientists in and outside Africa who have helped at various stages of my work in the Kibaran are too numerous to be listed here, but I continue to value their individual contributions. UNESCO has several times supported my participation in Geotraverses that were essential for a regional appraisal, and it assisted African scientists to take part in field conferences. The German Research Foundation (DFG) provided essential support for both field and laboratory work. As referees, David J. Vaughan and Michel Lécalle considerably helped to improve this paper.

References

- Aderca, B., 1957. Un cas de 'boudinage' à grande échelle: La mine de Rutongo au Rwanda. *Ann. Soc. Géol. Belg.*, 80B: 279–285.
- Barnes, J.W., 1961. The mineral resources of Uganda. *Bull. Geol. Surv. Uganda*, 4: 1–89.
- Baudet, D., Hanon, M., Lemonne, E. and Theunissen, K., 1988. Lithostratigraphie du domaine sédimentaire de la chaîne Kibarienne au Rwanda. *Ann. Soc. Géol. Belg.*, 112: 225–246.
- Baudin, B., Zigirabili, S. and Ziserman, A., 1984. Livre notice de la carte des gîtes minéraux du Rwanda (1:250.000). Ed. Univ. Natl. Rwanda, Butare, 163 pp.
- Bertossa, A., 1965. La pegmatite du Buranga. *Bull. Serv. Géol. Rwanda*, 2: 1–6.
- Bertossa, A., 1968. Inventaire des minéraux du Rwanda. *Bull. Serv. Géol. Rwanda*, 4: 25–46.
- Bowden, P. and Karche, J.-P., 1984. Mid plate A-type magmatism in the Niger–Nigeria anorogenic province: age variations and implications. In: J. Klerkx, and J. Michot (Editors), *African Geology*. Tervuren, pp. 167–177.
- Brinckmann, J., Lehmann, B. and Timm, F., 1994. Proterozoic gold mineralization in NW Burundi. In: W. Pohl (Editor) *Kibaran (Mid-Proterozoic) Metallogeny in Central and Southern Africa*. *Ore Geol. Rev.*, 9: 85–103.
- Cahen, L., Snelling, N.J., Delhal, J. and Vail, J.R., 1984. The geochronology and evolution of Africa. *Oxford Science Publ.*, Clarendon Press, Oxford, 512 pp.
- Campbell, A., Rye, D. and Petersen, U., 1984. A hydrogen and oxygen isotope study of the San Cristobal mine, Peru: implications of the role of water to rock ratio for the genesis of wolframite deposits. *Econ. Geol.*, 79: 1818–1832.
- Cauet, S. and Pohl, W., 1988. Lead isotope composition of sulfide minerals from Kibaran mineralisations in Rwanda. *IGCP 255; Newsletter/Bulletin*, 1: 11–13.
- Cerny, P., 1991. Fertile granites of Precambrian rare-element

- pegmatite fields: is geochemistry controlled by tectonic setting or source lithologies? *Precambrian Res.*, 51: 429–468.
- Costa, M., Ferrara, G., Sacchi, R. and Tonarini, S., 1992. Rb/Sr dating of the Upper Proterozoic basement of Zambesia, Mozambique. *Geol. Rundschau*, 81: 487–500.
- Daly, M.C., 1986. The intracratonic Irumide Belt of Zambia and its bearing on collision orogeny during the Proterozoic of Africa. In: M.P. Coward and A.C. Ries (Editors), *Collision Tectonics*. *Geol. Soc. Spec. Publ.*, 19: 321–328.
- Deblond, A., 1990. Late Kibaran layered igneous rocks from Eastern Burundi, a progress report. *IGCP 255; Newsletter/Bulletin*, 3: 9–17.
- Deblond, A., 1993. Géologie et pétrologie des massifs basiques et ultrabasiques de la ceinture Kabanga–Musongati au Burundi. Thèse Univ. Liège, Liège, 253 pp.
- De Magnée, I. and Aderca, B., 1960. Contribution à la connaissance du Tungsten-belt ruandais. *Acad. R. Sci. Outre-Mer*, 11/7: 1–56.
- Doe, B.R. and Zartman, R.E., 1979. Plumbotectonics. In: H.L. Barnes (Editor), *Geochemistry of Hydrothermal Ore Deposits*, 2nd ed., Wiley Interscience, New York, N.Y., pp. 22–70.
- Eriksson, K.A. and Chuck, R.G., 1985. Aulacogens: sedimentological and tectonic evolution and associated mineralization. In: K.H. Wolf (Editor), *Handbook of Strata-Bound and Stratiform Ore Deposits*, 12: 461–529.
- Fernandez-Alonso, M., Lavreau, J. and Klerkx, J., 1986. Geochemistry and geochronology of the Kibaran granites in Burundi, Central Africa: Implications for the Kibaran orogeny. *Chem. Geol.*, 57: 217–234.
- Franceschi, G., 1990. Primary gold mineralization in SW Rwanda. *IGCP 255; Newsletter/Bulletin*, 3: 19–25.
- Frisch, W., 1971. Die Zinn-Wolfram-Provinz in Rwanda (Zentral-Afrika) aus montangeologischer Sicht. *Erzmetall*, 24: 593–600.
- Frisch, W., 1975. Die Wolfram-Lagerstätte Gifurwe (Rwanda) und die Genese der zentralafrikanischen Reinit-Lagerstätten. *Jahrb. Geol. B.-A. Wien*, 118: 119–191.
- Gosse, R., 1992. The Kabanga Ni–(Co–Cu) sulphide deposit, Western Tanzania. *IGCP 255; Newsletter/Bulletin*, 4: 73–76.
- Guenther, M.A., Dulski, P., Lavreau, J., Lehmann, B., Möller, P. and Pohl, W., 1989. The Kibaran tin granites: Hydrothermal alteration versus plate tectonic setting. *IGCP 255; Newsletter/Bulletin*, 2: 21–27.
- Guenther, M.A., 1990. Flüssigkeitseinschlüsse und geologisches Umfeld zentralafrikanischer Sn-, W- und Au-Lagerstätten (Rwanda und Burundi). *Braunschweiger Geol. Paläontol. Diss.*, 11, 148 pp.
- Hartnady, Ch.J.H., 1991. About turn for supercontinents. *Nature*, 352: 476–478.
- Harris, N.B., Pearce, J.A. and Tindle, A.G., 1986. Geochemical characteristics of collision zone magmatism. In: M.P. Coward and A.C. Ries (Editors), *Collision Tectonics*. *Geol. Soc. London, Spec. Publ.*, 19: 67–81.
- Hutchison, C.S., 1982. The various granitoid series and their relationship to W and Sn mineralization. In: J.V. Hepworth and Yu Hong Zhang (Editors), *Tungsten Geology*. Jiangxi, China, ESCAP/RMRDC Bandung, Indonesia, pp. 87–114.
- Jacobs, J., Thomas, R.J. and Weber, K., 1993. Accretion and indentation tectonics at the southern edge of the Kapvaal craton during the Kibaran (Grenville) orogeny. *Geology*, 21: 203–206.
- Jedwab, J., 1987. Bref aperçu des minéraux de Pt, Pd et Ir trouvés par microsonde électronique dans les roches du massif ultrabasique de Musongati (Burundi). *Mus. R. Afrique Centr. Tervuren (Belgium), Rapp. Annu.* 1985–86: 83–87.
- Jedwab, J., 1992. A review of platinum occurrences in Zaire (except Shaba). *IGCP 255; Newsletter/Bulletin*, 4: 101–105.
- Jeffery, P.G., 1959. The geochemistry of tungsten, with special reference to the rocks of the Uganda Protectorate. *Geochim. Cosmochim. Acta*, 16: 278–295.
- Jourde, G. and Vialette, Y., 1980. La chaîne du Lurio (Nord Mozambique). *Int. Rep.*, BRGM-Orléans, 75 pp.
- Jung, D. and Meyer, F.-M., 1990. Les méta-basites du Rwanda. *IGCP 255; Newsletter/Bulletin*, 3: 37–50.
- Key, R.M., Charsley, T.J., Hackman, B.D., Wilkinson, A.F. and Rundle, C.C., 1989. Superimposed Upper Proterozoic collision-controlled orogenies in the Mozambique orogenic belt of Kenya. *Precambrian Res.*, 44: 197–225.
- Klerkx, J., Lavreau, J., Liégeois, J.-P. and Theuvsissen, K., 1984. Granitoïdes kibariens précoces et tectonique tangentielle au Burundi: magmatisme bimodal lié à une distension crustale. In: J. Klerkx and J. Michot (Editors), *African Geology*. Tervuren, pp. 29–46.
- Klerkx, J., Liégeois, J.-P., Lavreau, J. and Claessens, W., 1987. Crustal evolution of the northern Kibaran belt, eastern and central Africa. In: A. Kröner (Editor), *Proterozoic Lithospheric Evolution*. *Geodynamics Ser.*, Am. Geophys. Union, 17: 217–233.
- Kröner, A., 1984. Late Precambrian plate tectonics and orogeny: A need to redefine the term Pan-African. In: J. Klerkx and J. Michot (Editors), *African Geology*. Tervuren, pp. 23–28.
- Lavreau, J., 1984. Aperçu sur la géochronologie du Kibarien de l'Afrique Centrale. *UNESCO Geol. Developm. Newsl.* 3: 31–35.
- Lavreau, J. and Liégeois, J.-P., 1982. Granites à étain et granito-gneiss burundiens au Rwanda (région de Kibuye): âge et signification. *Ann. Soc. Géol. Belg.*, 105: 289–294.
- Lehmann, B. and Lavreau, J., 1988. Geochemistry of tin granites from Kivu (Zaire), Rwanda and Burundi. *IGCP 255; Newsletter/Bulletin*, 1: 43–46.
- Lowenstein, P.L., 1966. Progress report on studies of pegmatite and tin mineralization in south-west Ankole, Uganda. 16th Annu. Rep., Res. Inst. Afr. Geol., Univ. Leeds, pp. 34–36.
- Manning, D.A.C., 1984. Volatile control of tungsten partition-

- ing in granitic melt-vapour systems. IMM Trans., B93: 185–189.
- Maravic, H.V., Morteani, G. and Roethe, G. 1989. The cancrinite-syenite/carbonatite complex of Lueshe, Kivu/NE Zaire: petrographic and geochemical studies and its economic significance. *J. Afr. Earth Sci.*, 9(2): 341–356.
- McNaughton, N.J., 1987. Lead isotope systematics for Archean sulfide studies. *Geol. Dep. Univ. Ext., Univ. Western Australia Publ.*, 11: 181–188.
- Niyondezo, S., 1984. Les ressources minérales du Kibarien au Burundi. UNESCO Geol. Developm. Newsl., 3: 37–41.
- Ntungicimpaye, A., 1984. Le magmatisme basique dans le Burundien de l'Ouest du Burundi. UNESCO Geol. Developm. Newsl., 3: 13–21.
- Ntungicimpaye, A. and Tack, L., 1992. Les métavolcanites intermédiaires à acides kibariennes du NW du Burundi. IGCP 255; Newsletter/Bulletin, 4: 45–50.
- Pargeter, R.C. 1952. The geology of bismuth ores in Uganda. *Proc. 5th Int. Geol. Congr. East Africa*.
- Pargeter, R.C., 1956. The Ruhiza ferberite deposit, Kigezi. *Rec. Geol. Surv. Uganda*, 1954: 27–46.
- Pinna, P., Jourde, G., Calvez, J.Y., Mroz, J.P. and Marquez, J.M., 1993. The Mozambique Belt in northern Mozambique: Late Proterozoic (1100–8500 Ma) crustal growth and tectogenesis, and superimposed Pan-African (800–550 Ma) tectonism. *Precambrian Res.*, 62: 1–59.
- Plumb, K.A., 1991. New Precambrian time scale. *Episodes*, 14: 139–140.
- Pohl, W., 1975. Géologie de la mine de Bugarama et de ses environs (Rwanda, Afrique). *Bull. Serv. Geol. Rwanda*, 8: 13–42.
- Pohl, W., 1977. Structural control of tin and tungsten mineralization in Rwanda, Africa. *Berg. Huettenmaenn. Monatsh.*, 122: 59–63.
- Pohl, W., 1978. Die tektonische Kontrolle der Zinngänge von Rutongo, Rwanda (Afrika). *Mitt. Österr. Geol. Ges.*, 68: 89–107.
- Pohl, W., 1986. Geotraverse Rwanda 1986: Preliminary results of metallogenetic studies. UNESCO Geol. Econ. Developm. Newsl., 5: 93–100.
- Pohl, W., 1987a. Metallogeny of the northeastern Kibaran belt, Central Africa. *Geol. J.*, 22: 103–119.
- Pohl, W., 1987b. Manganiferous garnetite bands in Kibaran meta-volcanosedimentary rocks at Kibuye, Rwanda: Exhalites of metallogenic significance? In: G. Matheis and H. Schandelmeier (Editors), *Current Research in African Earth Sciences*. Balkema, Rotterdam, pp. 349–351.
- Pohl, W., 1992. Kibaran evolution and metallogeny in Central Africa: a synthesis at the end of IGCP-Project 255. IGCP 255; Newsletter/Bulletin, 4: 1–8.
- Pohl, W. and Hadoto, D.P.M., 1990. Granite-related Kibaran gold mineralization at Mashonga, Bushenyi District (SW-Uganda). IGCP 255 Newsletter/Bulletin, 3: 61–67.
- Pohl, W. and Guenther, M.A., 1991. The origin of Kibaran (late Mid-Proterozoic) tin, tungsten and gold quartz vein deposits in Central Africa. A fluid inclusions study. *Mineral. Deposita*, 26: 51–59.
- Prochaska, W. and Rulinda, G., 1989. The talc deposit of Kibanda (Rwanda). IGCP 255 Newsletter/Bulletin, 2: 65–72.
- Radulescu, J., 1982. Mineralization in the Karagwe-Ankolean System of East Africa/ Burundi. In: *The Development Potential of Precambrian Mineral Deposits*. UN Dep. Tech. Coop. Developm., pp. 217–225.
- Reedman, A.J., 1967. The geological environment and genesis of the tungsten deposits of Kigezi district, south-western Uganda. Ph.D. thesis abstract, 1967, 11th Annu. Rep. Res. Inst. African Geol., Univ. Leeds, 38 pp.
- Robert, M., 1931. Carte géologique du Katanga 1:1,000,00. *Nouv. Mém. Soc. Belge Géol. Paléontol. Hydrogr.*, 5: 1–14.
- Roedder, E., 1984. Fluid inclusions. *Reviews in Mineralogy*, 12. Mineral. Soc. Am., Washington DC, 644 pp.
- Routhier, P. 1980. Où sont les métaux pour l'avenir? *Mém. BRGM Orléans* 185, 408 pp.
- Rumvegeri, B.T. and Katararwa, J.-B., 1990. Géochimie des granitoïdes kibariens du Kivu (Est-Zaire) et du Rwanda: implications géodynamiques. *C.R. Acad. Sci. Paris*, 311/II: 959–963.
- Sacchi, R., Marques, J., Costa, M. and Casati, C., 1984. Kibaran events in the southernmost Mozambique Belt. *Precambrian Res.*, 25: 141–159.
- Tack, L., 1990. Late Kibaran structural evolution in Burundi. IGCP 255 Newsletter/Bulletin, 3: 77–79.
- Tack, L. and Deblond, A., 1990. Intrusive character of the late Kibaran magmatism in Burundi. IGCP 255 Newsletter/Bulletin, 3: 81–87.
- Tack, L., De Paepe, P., Liégeois, J.P., Nimpagaritse, G., Ntungicimpaye, A. and Midende, G., 1990. Late Kibaran magmatism in Burundi. *J. Afr. Earth Sci.*, 10: 733–738.
- Tack, L., Duchesne, J.C., Liégeois, J.P. and Deblond, A., 1994. Two successive mantle-derived A-type granitoids in Burundi: Kibaran late-orogenic extensional collapse and lateral shear along the edge of the Tanzanian craton. *Precambrian Res.*,
- Tassinari, C.C.G., Siga, O. and Barreto, L.M.B., 1989. The geochronology of Kibaran belts and associated mineralizations in Brazil: a review. IGCP 255 Newsletter/Bulletin, 2: 77–83.
- Theunissen, K., 1989. On the Rusizian basement rise in the Kibara belt of Northern Lake Tanganyika. IGCP 255; Newsletter/Bulletin, 2: 85–92.
- Thomas, R.J., Agenbacht, A.L.D., Cornell, D.H. and Moore, J.M., 1994. The Kibaran of Southern Africa: tectonic evolution and metallogeny. In: W. Pohl (Editor) *Kibaran (Mid-Proterozoic) Metallogeny in Central and Southern Africa*. *Ore Geol. Rev.*, 9: 131–160.
- Tissot, F., Swager, C., Berg, R., Van Straaten, H.P. and Ingotatow, A., 1982. Mineralization in the Karagwe-Ankolean System of North-West Tanzania. In: *The Development Potential of Precambrian Mineral Deposits*. UN Dep. Techn. Coop. Developm., pp. 205–215.

- Van Straaten, H.P., 1984. Contributions to the geology of the Kibaran belt in Northwest Tanzania. UNESCO Geol. Developm. Newsl., 3: 59–68.
- Van Wambecke, L., 1977. The Karonge rare earth deposits, Republic of Burundi. *Mineralium Deposita*, 12: 373–380.
- Varlamoff, N., 1969. Transitions entre les filons de quartz et les pegmatites stannifères de la région de Musha–N'tunga (Ruanda). *Ann. Soc. Géol. Belge*, 92: 193–213.
- Varlamoff, N., 1972. Central and West African rare metal granitic pegmatites, related aplites, quartz veins and mineral deposits. *Mineralium Deposita*, 7: 202–216.
- Varlamoff, N., 1975. Classification des gisements d'étain. *Acad. R. Sci. Outre-Mer*, 19(5): 1–63.
- Von Knorring, O., 1969. A note on the phosphate mineralization at the Buranga pegmatite, Rwanda. *Bull. Serv. Géol. Rwanda*, 5: 42–45.
- Windley, B.F., 1984. *The Evolving Continents*, 2nd ed. Wiley, Chichester, 399 pp.
- Ziegler, P.A., 1987. Late-Cretaceous and Cenozoic intra-plate compressional deformations in the Alpine foreland — a geodynamic model. *Tectonophysics*, 137: 389–420.
- Ziserman, A., Zigirabili, J., Petricec, V. and Baudin, B., 1983. Données sur la métallogénie du Rwanda. Enseignements tirés de la carte des gîtes minéraux. *Chron. Rech. Min.* 471: 31–40.