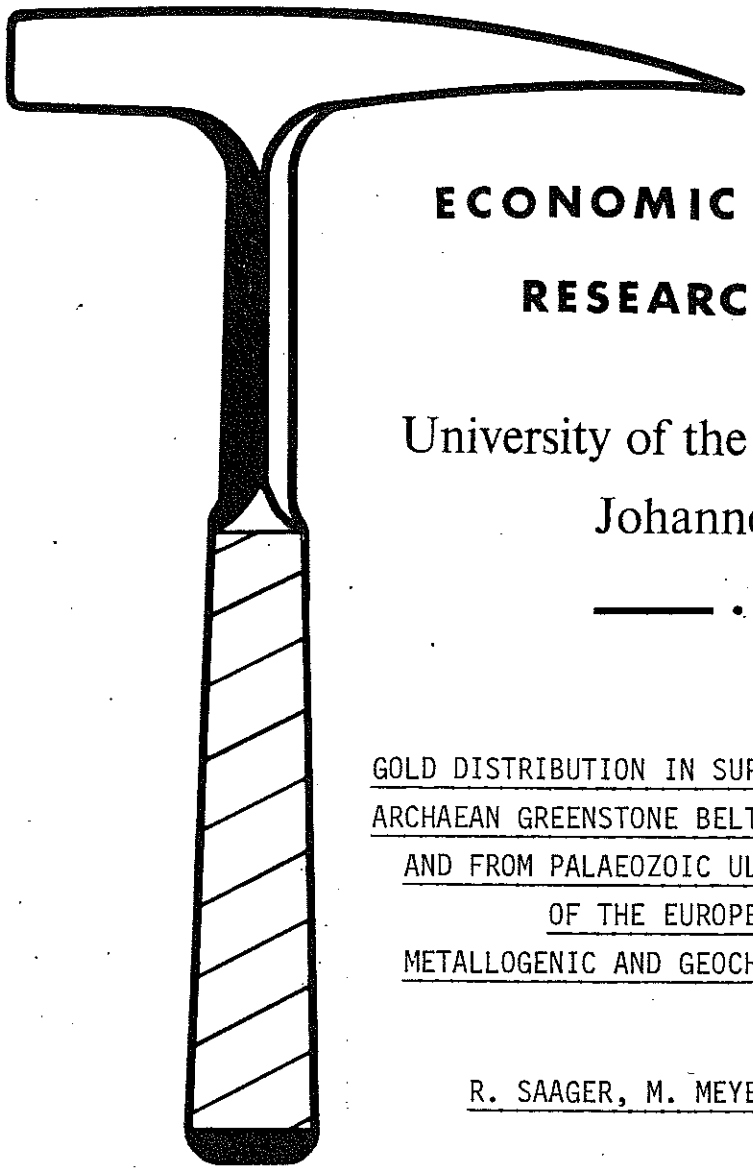

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GOLD DISTRIBUTION IN SUPRACRUSTAL ROCKS FROM ARCHAEAN GREENSTONE BELTS OF SOUTHERN AFRICA AND FROM PALAEOZOIC ULTRAMAFIC COMPLEXES OF THE EUROPEAN ALPS : METALLOGENIC AND GEOCHEMICAL IMPLICATIONS

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by

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ABSTRACT

Many Archaean gold veins are thought to have formed as a result of metamorphic secretion processes, with the source of gold being "gold enriched" supracrustal rocks. However, in many gold provinces no elevated gold concentrations can be detected in the proposed "source beds".

In this study, 98 samples of mafic and ultramafic rocks, and 32 samples of ferruginous chemical sediments from South African greenstone belts and from the Belingwe greenstone belt in Zimbabwe were analysed for their Mg, Al, Cr, Fe, Co, Ni, and Au contents. The analyses were carried out by means of XRF, AAS, and NAA, and the data evaluated by various statistical methods. The gold values of the volcanics range from 0,1 to 372 ppb (with the mean content lying at 10,8 ppb) and the gold values of the sediments vary from 0,5 to 667 ppb (the mean content being 129,9 ppb). As a comparative population, 56 samples from Palaeozoic European ultramafic complexes were investigated for the same suite of elements. The gold values range from 0,1 to 25 ppb and possess a mean content of 2,0 ppb.

For the samples from southern Africa, statistical analyses indicate complete lack of relationship between gold and rock-forming minerals. Significant differences of gold abundance were observed in samples originating from geologically and geochemically similar greenstone terranes. These differences may be a result of heterogeneous gold abundance in the upper mantle. Differentiation trends of gold in the ultramafic-to-mafic range were not detected in this study. Ferruginous chemical sediments of the Algoma-type were found to contain more gold than younger ones of the Superior-type.

For volcanic rocks, the analytical data suggests that the principal gold carriers are accessory sulphides and, to a lesser extent, intergranular gold in particulate form. Rock-forming minerals appear to contribute less than 0,5 ppb to the gold in these rocks. Most supracrustal rocks contain sulphides and, thus, contain gold readily accessible to leaching metamorphic hydrothermal solutions which may deposit gold in nearby dilatant zones. It is proposed that in greenstone terranes metamorphic overprint constitutes the key factor determining the formation of gold veins. The conspicuous abundance of such gold mineralizations in Archaean greenstone terranes is considered to be a result of the unique evolution of these terranes.

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I. INTRODUCTION

The world-wide acceleration of exploration activities for gold, which accompanied the recent rises in the gold price, clearly exposed the necessity to improve our knowledge of the geochemistry of gold, and to define more precisely the parameters responsible for the deposition of economic lode gold concentrations. This development followed closely the introduction of rapid, highly sensitive gold assaying methods and, for the past decade, resulted in the publication of an increasing number of gold analyses of geologic material.

In line with the demand for an enhanced understanding of the gold geochemistry the objectives of the analytical studies were broadly as follows :

(i) to investigate the background levels of gold in various types of igneous rocks, and to assess the abundances of gold in rock-forming minerals;

(ii) to ascertain whether gold follows a differentiation trend during the crystallization of magmas, e.g. to formulate the laws controlling the distribution of gold in igneous rocks;

(iii) to investigate the possible occurrence of regional variations in the gold content of igneous rocks which could indicate inhomogeneities in the earth's mantle, or differences in the tectonic setting of the rocks studied; and finally,

(iv) to establish metallogenetic models for the formation of hydrothermal gold deposits and to recognize the ultimate source of the gold in these deposits.

For a comprehensive bibliography of the geochemistry of gold the reader is referred to the compilations of Cooper (1971) and Crocket (in Wedepohl, 1974), and to the review papers by Gottesmann (1976), Foster (1977), and Boyle (1979). Here are mentioned only some of the publications dealing particularly with metallogenetic aspects (Fersman, 1939; Gallagher, 1940; Shilin, 1968; Mantai and Brownlow, 1967; Viljoen *et al.*, 1969; Stephenson and Ehmann, 1971; Gottfried *et al.*, 1972; Tilling *et al.*, 1973; Weber and Stephenson, 1973; Seeland, 1973; Henley, 1973; Pyke, 1975; Keays and Kirkland, 1972; Keays and Scott, 1976; Kwong and Crocket, 1978; Sawkins *et al.*, 1979).

The large amount of data gathered by various workers, and the often differing results obtained, have led to the formulation of a number of frequently contradictory goechemical models for the abundance of gold in geologic material. Data interpretations have been,and still are, hampered by the fact that very little is known as to how gold is distributed in rocks and rock-forming minerals. Uncertainty exists as to whether it occurs as minute intergranular particles of native gold, or whether it occurs as submicroscopic contaminants in minerals and/or in solid solution with certain minerals and mineral groups.

As regards differentiation, many workers, including Shcherbakov (1965, 1967), Jones (1969), Shilin (1968), Gottfried *et al.* (1972), and Tilling *et al.* (1973), observed a general decrease in the gold content from mafic to felsic calc-alkalic igneous rocks.

For ultramafic rocks, high gold values have been reported from the Urals (5-500 ppb, Borisenko *et al.*, 1972), from Manitoba (17-19 ppb, Stephenson and Ehmann, 1971), from the USSR (1,4-63 ppb, Moiseenko *et al.*, 1971), and from South Africa (10-20 ppb, Viljoen *et al.*, 1969).

A preferred association of higher gold values with mafic and ultramafic rocks seems to be confirmed by the close spatial relationship of many Archaean gold deposits with rocks of this type as well as the occurrence of relatively high gold contents in many ores of mafic and ultramafic associations (e.g. the Merensky Reef of the Bushveld Igneous Complex, Sudbury, and many other Ni-Co deposits in Canada and Western Australia, Stanton, 1972).

Volcanic rocks usually show a higher variability of their gold contents than plutonic rocks (Tilling *et al.*, 1973), and there is also a general tendency for mafic varieties to have more gold than felsic varieties (Gottfried *et al.*, 1972). Large discrepancies in gold abundance have been found in low-K_2O oceanic tholeiites (De Grazia and Haskin, 1964; Gottfried *et al.*, 1972; Laul *et al.*, 1972; Keays and Scott , 1976). Tilling *et al.* (1973) cite the low gold values in oceanic tholeiites of the Mid-Atlantic Ridge and East Pacific Rise as a good example of regional or provincial differences in gold abundance for basaltic rocks.

Lack of concentration of gold in residual silicate melts during differentiation of magmas has been discussed by Tilling *et al.* (1973). Sadowski *et al.* (1971) found that rock types that crystallized early in a differentiated suite possess higher gold values than the late precipitates. However, Gottfried and Greenland (1972), who studied variations of iridium and gold in oceanic and continental basalts, concluded that gold has an inert behaviour during differentiation. Nearly uniform gold contents in the various rock types of the extremely fractionated Skaergaard and Zlatogosk intrusions were reported by Vincent and Crocket (1960) and Voskresenskaya and Zvereva (1968).

Available values of gold content in sedimentary rocks indicate a wide scatter within distinct types of sedimentary rocks. Kwong and Crocket (1978) found that in the Archaean greenstone assemblage of the Kakagi Lake area in Northwestern Ontario the sedimentary rocks (tuffwackes,cherts, argillites and siltstones) showed, on average, significantly lower gold values than the mafic volcanics. From 44 samples of interflow sedimentary rocks from within the komatiite sequence of Kambalda, Bavinton and Keays (1978) report a mean gold content of 142,6 ppb. These authors assumed that these sediments, which consist predominantly of iron-sulphide-bearing siliceous material, represent a mixture of volcanoclastic material associated with the extrusion of the

ultramafic-mafic lavas, and chemical-sedimentary material. Saager and Muff (1978) reported an even higher mean gold content (204 ppb) for 19 samples of volcanogenic oxidic banded iron-formation collected from the Archaean greenstone belts of South Africa.

Investigations on the gold contents of minerals indicate that the average gold concentration decreases from magnetite and ferromagnesian silicates to feldspar and quartz (Shcherbakov, 1967; Davletov, 1970; Gottfried *et al.*, 1972). Shcherbakov and Perezhogin (1963) suggested that during differentiation gold, due to its high electronegativity, preferentially accumulates in the gaseous phases that separate from magmas, and that those silicates which possess a minimum amount of ionic character, i.e. nesosilicates and isosilicates, are the most likely to contain chemically bound gold. Such a model partly explains the relatively higher average gold contents of olivines and pyroxenes and, correspondingly, the higher gold content of mafic and ultramafic rocks. Vincent and Crocket (1960) observed in the Skaergaard intrusion that gold is fairly evenly distributed among the predominant silicate and oxide minerals and that the metal probably is present as occasional atoms in defect structures of the crystal lattice of the rock-forming minerals. These authors also stressed the close similarity of the ionic radii of copper and gold and suggested that traces of gold may enter copper sulphide minerals. In the Great Lake tholeiitic dolerite Rowe (1969) reported a positive correlation between gold and copper and the mafic index with gold reaching a ceiling at 12 ppb at a mafic index of 70. It thus seems probable that depletion of gold in magmas can take place when saturation with respect to copper sulphides is reached.

Keays and Scott (1976) concluded that in the case of a crystallizing basaltic melt the bulk of the gold is concentrated in early separated oxide and silicate phases and in sulphide phases. Their suggestion is supported by high correlations between Cr, Ni, MgO, and Au contents.

Moiseenko *et al.* (1972), Sawkins *et al.* (1979) and others, concluded from fluid inclusions studies that hydrothermal gold deposits generally formed below 300°C and that maximum gold deposition probably took place in the temperature range of 200-300°C. With respect to the solubility and transport of gold, experimental data indicate that gold is transported as complex ionic species and is deposited following changes of P, T, pH, Eh, fO_2, and total sulphur content (Seward, 1973; Henley, 1973). Boyle *et al.* (1975) and Pyke (1976) also stresseed the importance of carbonatization, sericitization and albitization in many areas where hydrothermal gold deposits occur. Such alteration processes may thus be important factors for the release of gold, especially from ultramafic rocks.

The close spatial relationship between hydrothermal gold deposits and mafic to ultramafic metavolcanics in Precambrian greenstone belts has been known for a long time and is described in a number of publications from many shield areas of the world (Lightfoot, 1930; Macgregor, 1951; Boyle, 1961; Viljoen *et al.*, 1969; Saager, 1973; Weber and Stephenson, 1973; Pyke, 1975, 1976; Anhaeusser, 1976; Keays and Davidson, 1976). Precambrian greenstone belts are thus also referred to as "gold belts". In many cases, the gold occurrences are situated within the mafic or ultramafic members of the volcanic piles, although gold veins also occur in intermediate to acid volcanics and accompanying sedimentary rocks. The occurrence of the gold veins is to a large extent, controlled by the competency of their host rocks (Viljoen *et al.*, 1969).

Some investigators of early Precambrian gold lode deposits have maintained that the gold was derived from granitic plutons which intruded the greenstone terranes (Emmons, 1937; De Villiers, 1957; George, 1967; Jones, 1948). Boyle (1961), however, introduced the source-bed concept to explain the genesis of Precambrian epigenetic gold occurrences whereby the gold was derived by lateral secretion from the surrounding volcanic rocks. Viljoen *et al.*, (1969) suggested that the ultimate source of the gold lies in "primitive" mafic to ultramafic volcanics from which it was mobilized during regional metamorphism caused by the intrusion of Na-rich granitic plutons. Following this event quartz-gold veins developed in adjacent competent rocks.

Many workers suggested that the "source beds" should contain above normal gold concentrations (Viljoen *et al.*, 1969; Pyke, 1975; Stephenson and Ehmann, 1971). However, in some gold provinces, no elevated gold contents could be detected in the proposed "source rocks". Tilling *et al.* (1973) concluded that the background gold levels of various rock types are of no diagnostic importance for recognizing areas favourable for gold mineralization. In a study of the Archaean Kakagi Lake area in Canada, Kwong and Crocket (1978) stated that "the potential for mineralization is only weakly dependent on gold abundance levels", and Keays and Scott (1976) stressed that the suitability of any rock as a source for gold mineralization is largely determined by the presence of "excess" or "reactive" gold in the source rock, e.g. by the occurrence of gold in sulphides, mesostasis phases, and on grain boundaries.

Fripp (1976a, b), commenting on the formation of stratiform Archaean banded iron-formations, proposed that the sulphides, carbonates, cherts, and gold are syngenetic submarine sedimentary accumulations which precipitated out from geothermal brines during the final stages of volcanic events.

The present study was initiated to contribute to the understanding of the genesis of vein-type gold deposits in Archaean volcanic regions. A further objective was to gather more data on the distribution of gold in Archaean volcanic rocks in southern Africa, a region which constitutes an important gold province, but from which only a few gold analyses of rock samples have been published to date (Viljoen *et al.*, 1969; Anhaeusser *et al.*, 1975; Saager and Muff, in press). The geologic setting, geochemistry, and petrology of the greenstone regions sampled (i.e. the Barberton Mountain Land, the Sutherland Range, and the Pietersburg Belt in South Africa, as well as the Belingwe Belt in Zimbabwe (Fig. 1), are well-studied and, thus, constitute ideal terranes for testing geochemical criteria pertinent to the metallogenesis of gold veins.

In order to compare the gold contents of the Archaean volcanic rocks with much younger ultramafic and mafic rocks having a different petrogenetic history and originating from different tectonic environments, a number of analyses from peridotites and amphibolites of the Palaeozoic Finero Complex in the Southern Alps of Switzerland/Northern Italy, and from serpentinized dunites, serpentinites and amphibolites of the Palaeozoic Hochgrössen and Kraubath Massifs of the Eastern Alps, Austria, have been incorporated in this study (Fig. 2).

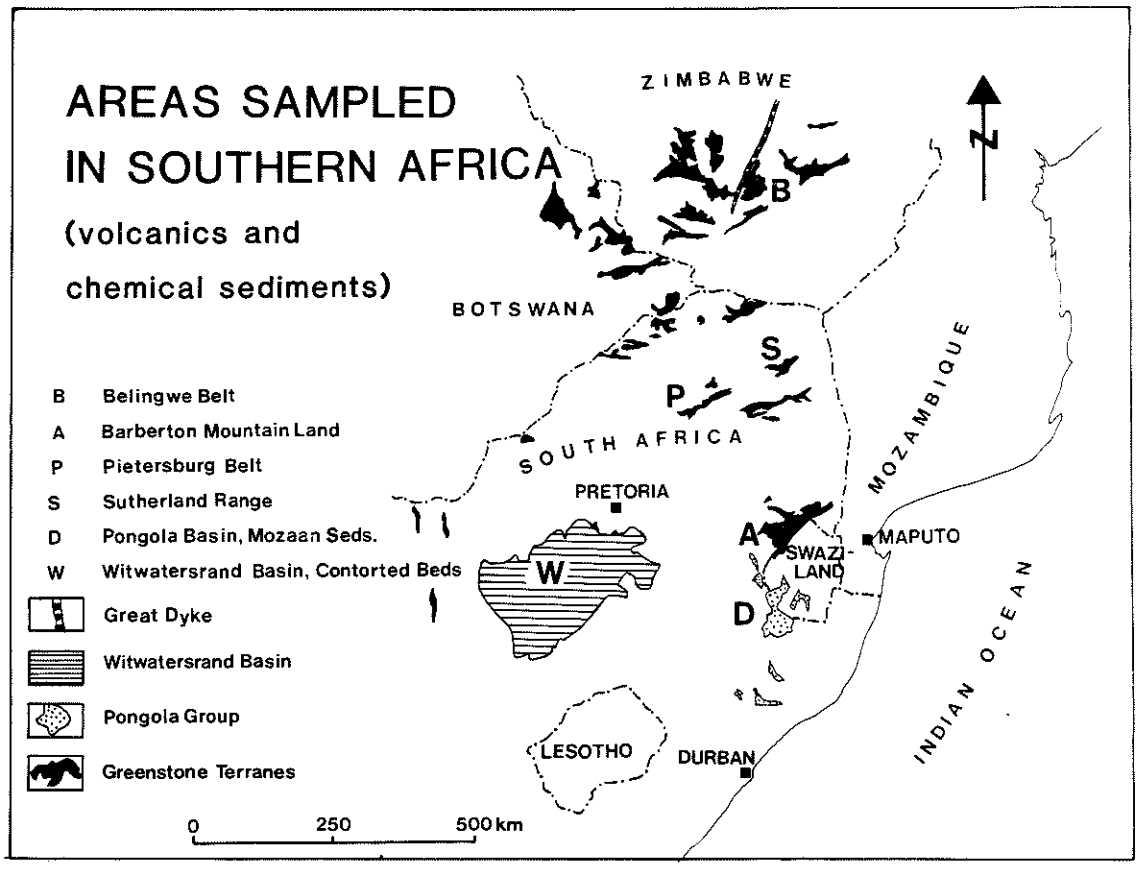

Figure 1 : *Sketch map of southern Africa showing areas sampled in this study.*

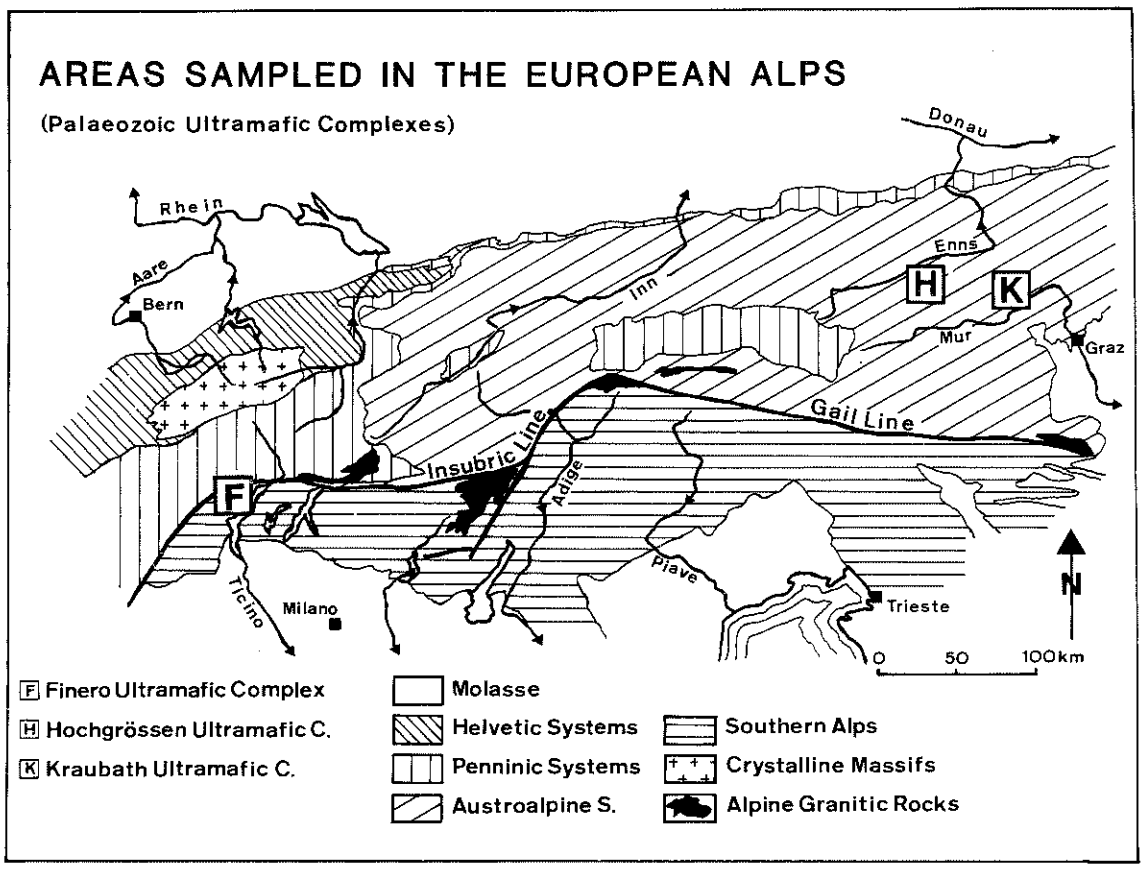

Figure 2 : *Geologic sketch map of the central and eastern Alps showing the localities of the three ultramafic complexes sampled.*

II. GEOLOGICAL SETTING OF SAMPLED AREAS AND SAMPLE DESCRIPTION

A. Barberton Mountain Land

Twelve samples were collected from the Archaean Barberton greenstone belt in the Eastern Transvaal of South Afica (Fig. 1). The rocks consist of peridotitic and basaltic komatiites from the type area of the Komati Formation (Tjakastad Subgroup) in the Komati River Valley. This area was selected as it provides the best preserved and least altered extrusive komatiitic and basaltic lavas in the Barberton district. No gold was produced in the environs of the sampling area — all the important gold mines of the Barberton Mountain Land lying some 20 km to the northeast. The geology, petrography and geochemistry of the volcanic rocks in the area sampled were described by Viljoen and Viljoen (1969a). These authors divided the volcanic sequence into a predominantly komatiitic Lower Ultramafic Unit (the Tjakastad Subgroup) and an overlying essentially tholeiitic and calc-alkalic Mafic-to-Felsic Unit (the Geluk Subgroup) . The volcanic sequence has an age of between 3500 - 3800 Ma old (Jahn and Shih, 1974; Saager and Köppel, 1976).

In the area sampled ultramafic lava flows alternate with basalts. Viljoen and Viljoen (1979) used a MgO vs Al_2O_3 plot to group the various komatiitic rocks into peridotitic, pyroxenitic and basaltic komatiites and to demarcate the komatiites from the tholeiites (Fig. 3). From this plot it is apparent that the South African data, where compared with the data from the considerably younger Belingwe Belt of Zimbabwe, exhibit an Al_2O_3 depleted trend.

The rocks in the area sampled have been subjected to greenschist facies metamorphism that caused hydration and, in places, also carbonatization. All the peridotitic komatiites examined are extensively serpentinized, although macroscopically they still look remarkably fresh. The modal percentage of olivine in the peridotitic komatiites varies from 45% to 60% (Viljoen and Viljoen, 1969b). Microscopically olivine is only rarely preserved and has been altered to antigorite associated with tremolite, chlorite, magnetite, rare ilmenite, chromite and a few grains of pyrite. The basaltic komatiites are also altered and are composed of stubby laths or slender needles of amphiboles, oligoclase, chlorite, zoisite, epidote, secondary carbonate, and quartz. Opaque minerals are rare and comprise magnetite, ilmenite, pyrrhotite, pyrite, arsenopyrite, and chalcopyrite. With the exception of very few exsolution flames of pentlandite in pyrrhotite, no nickel sulphides were observed.

B. Pietersburg Belt

Thirty-five samples were collected from volcanic rocks found near Eersteling in the Pietersburg Belt which is situated some 300 km north of Johannesburg in the northern Transvaal (Fig. 1). Recent studies carried out in the Pietersburg Belt include those of Grobler (1972). Saager and Muff (1978, and in press), Muff and Saager (1979), and Brandl (1979). Grobler (1972) correlated the volcanic assemblages at the base of the greenstone sequence with the Lower Ultramafic Unit and the Mafic-to-Felsic Unit of the Barberton Mountain Land. The lower unit, containing ultramafic and mafic, often komatiitic, extrusives was named the Mathiba Formation by Brandl (1979). It is separated from the overlying Eersteling Formation by the Ysterberg Formation, which probably represents an assemblage of volcanogenic oxidic banded iron-formations, ferruginous cherts, and cherts. The Eersteling Formation is made up of massive, generally tholeiitic, basalts which contain a number of broadly contemporaneous orthopyroxenite-peridotite sills, and minor intercalations of intermediate-to-acid lavas and banded ferruginous cherts.

The Pietersburg greenstone belt underwent regional greenschist metamorphism leading to the complete serpentinization of the ultramafic rocks. No olivine relics were observed in the samples investigated, but the primary presence of olivine is well-documented by secondary magnetite, ilmenite, and "mesh textures" outlining the shapes of the altered olivine crystals.

The tholeiitic metabasalts of the Eersteling Formation consist of actinolite, chlorite, epidote and minor amounts of feldspar and quartz. Carbonate minerals occur very rarely. A few scattered grains of pyrite, rutile, and magnetite were the only ore minerals observed in these rocks.

Five samples were obtained from the orthopyroxenite-peridotite sills which carry relatively high amounts of accessory opaque minerals. Chromite grains in these samples are peripherally replaced by magnetite which is the most abundant ore mineral. Other opaques include ilmenite, pyrrhotite, chalcopyrite and pyrite. Of particular interest is the occasional appearance of large, irregular,grains of pentlandite (0,5 mm diameter) closely intergrown with small amounts of pyrrhotite and chalcopyrite. The presence of pentlandite probably accounts for the high $Ni_{sulf}/Ni_{tot.}$ ratios observed in these rocks (Table 2).

Samples of banded ferruginous chert and oxidic banded iron-formation from the Ysterberg and Eersteling formations were also analysed for their gold contents and constitute the bulk of the chemical sediments investigated. Polished sections of banded iron-formation revealed the presence of hematite and supergene, colloform, iron-hydroxides which replace hematite. The microscopic texture of the hematite shows that it formed from martitization of magnetite and magnetite and hematite intergrowths were observed in ferruginous chert samples from the Eersteling Formation. Pyrite and pyrrhotite are rare and generally occur in the banded iron-formation chert.

C. Belingwe Greenstone Belt

Powdered material from 51 samples from the Belingwe greenstone belt in Zimbabwe was made available to the authors by Dr. E.A. Nisbet (Cambridge University) and the petrography and major element geochemistry of all the samples examined have been discussed by Nisbet *et al.* (1977).

The Belingwe greenstone belt lies some 450 km south of Salisbury, just east of the Great Dyke (Fig. 1). The geology, petrology and geochemistry of the belt have been studied by Macgregor (1947), Laubscher (1963),

Bickle *et al.* (1975), Martin (1975) and Nisbet *et al.* (1977). The Belingwe Belt is made up of two distinct suites of greenstone rocks. The basal sediments of the Upper Greenstone assemblage rest unconformably on folded and eroded remnants of the Lower Greenstones and, in parts, on foliated granitic basement. Jahn and Condie (1976) and Hawkesworth *et al.* (1977) report a Rb-Sr age of 2760 Ma for volcanics in the Upper Greenstone assemblage. Moorbath (pers. comm. in Nisbet *et al.* 1977) suggested an age of 3500 Ma for the foliated tonalitic basement granite.

The Belingwe samples consist of peridotitic komatiites, basaltic komatiites and high magnesian tholeiites which were collected from lava flows of the Upper Greenstone assemblage along the Ngesi River. According to Nisbet *et al.* (1977), the samples are partly altered and the ultramafic rocks are serpentinized. The Upper Greenstones are highly altered. No gold deposits are known from the Upper Greenstones.

D. Finero Ultramafic Complex

Forty two samples were collected in the Finero Complex, a lens of interlayered peridotite and amphibolite, situated in the Ivrea-Verbane Zone at the border between Switzerland and Northern Italy some 100 km northwest of Milano (Fig. 2). The Finero Complex has been studied by a number of workers and has been interpreted as (i) an alpine-type peridotite which intruded into basement rocks (Vogt, 1962); (ii) an upthrusted slice of upper mantle (Lensch, 1968; Ernst, 1977); or (iii) a layered intrusion formed by fractional crystallization of a hydrous magma (Cawthorn, 1975). Recent work on the sulphide mineralogy of rocks from the Finero Complex has been carried out by Oberthür (1978) who also provided some of the samples used in the present study. In addition, Meyer (1978) investigated the distribution of Au and Ir in the peridotite and amphibolite of the complex.

The Finero Complex has been subjected to multiple stages of metamorphism ranging from granulite to greenschist facies and which also caused serpentinization of certain portions of the complex. The granulite facies metamorphic event has been dated at 475 Ma whereas the greenschist facies metamorphism has an age of 15 Ma (Hunziker, 1974; Steck and Tièche, 1976).

Petrographically, the Finero Complex consists of phlogopite-peridotite, hornblende-peridotite and amphibolite. The phlogopite-peridotite samples are made up of olivine, diopside, enstatite, phlogopite, dark red-brown spinel, magnetite and hornblende. The hornblende-peridotite samples are coarse-grained and contain olivine, hornblende, diopside, enstatite, spinel and sulphide minerals, but they are devoid of phlogopite. The hornblende-peridotite has been partly serpentinized — some rocks being considerably recrystallized while others show thin cross-cutting veins of serpentine. Enrichments of Fe-, Cu- and Ni-sulphides are occasionally associated with the serpentinization. The metagabbro is banded, with plagioclase-rich light coloured layers and hornblende- and augite-rich dark coloured layers. Accessory minerals include spinel and rare grains of pyrrhotite, chalcopyrite and pentlandite.

E. Hockgrössen and Kraubath Ultramafic Bodies

Fourteen samples examined in the study originate from two metamorphosed ultramafic-mafic complexes, the Hochgrössen and Kraubath massifs situated in the Eastern Alps of Austria (Fig. 2). El Ageed (1979) and El Ageed *et al.* (1980) proposed that the two bodies are parts of a dismembered obducted ophiolite complex. The Kraubath ultramafics are believed to be the refractory residuals of this complex whereas the Hochgrössen rocks are considered to be a related cumulate ultramafic-mafic association. The rocks of both bodies are altered by serpentinization, metamorphism and tectonic deformation. No age data are available from either of the massifs but El Ageed (1979) has suggested a Palaeozoic age for their emplacement.

Samples from both massifs were made available for this study by Dr. A.I. El Ageed (Khartoum). The material from Kraubath consists of dunites, harzburgites, and bronzitites, all of which show advanced stages of serpentinization. Olivine and pyroxene have been altered to mesh-textured serpentine minerals, probably of the lizardite type. Opaque minerals include chromium spinel, secondary magnetite with subordinate pentlandite, pyrrhotite, chalcopyrite, and heazlewoodite. The Hochgrössen samples originate from serpentinized dunites and talc-carbonate rocks as well as from garnet-amphibolites. Altered olivine, antigorite, chlorite, and diopside are the main constituents of the dunites. The main opaque minerals in the dunites are chromite, magnetite, and Ni- and Cu-sulphides together with pyrrhotite. The talc-carbonate rocks are products of localized CO_2-metasomatism. Minerals present include magnesite, talc, tremolite, and antigorite. The garnet-amphibolites contain hornblende, plagioclase, garnet, idocrase, epidote, chlorite, and micas and the opaques include ilmenite and rutile with rare grains of pyrite, pyrrhotite, and chalcopyrite.

III. ANALYTICAL PROCEDURE

The samples from the South African greenstone belts, the Finero Complex, and the Hochgrössen and Kraubath massifs were analyzed for their major element contents by X-ray fluorescence. The results for Mg, Al, and Fe are given in Tables 1, 2, 4, and 5.

The samples from the Belingwe greenstone belt in Zimbabwe are duplicates of those discussed in the papers of Nisbet *et al.* (1975), and the Mg, Al, and Fe values given in Table 3 were obtained from these two publications.

The Cu contents of the samples from the Finero Complex (Table 4) and the $Ni_{sulf.}$ contents (e.g. the amount of Ni occurring in the sulphide phases of the samples) of the samples from southern Africa (Tables 1, 2, 3) were determined by flameless atomic absorption spectrometry. To determine $Ni_{sulf.}$, a sulphide selective leach was employed, using potassium chlorate-hydrochloric acid (as suggested by Olade and Fletcher, 1974).

The Cr, Ni, Co, and Au concentrations of all investigated samples were determined by instrumental epithermal neutron activation analysis (IENAA), and the results of the analyses are given in Tables 1 - 5. The

TABLE 1

GEOCHEMICAL DATA FOR SAMPLES COLLECTED AT THE TYPE LOCALITY OF THE KOMATI FORMATION IN THE BARBERTON MOUNTAIN LAND

Sample	MgO %	Al_2O_3 %	Cr ppm	Fe_2O_3+ %	Co ppm	$Ni_{tot.}$ ppm	$Ni_{sulf.}$ ppm	Au ppb	Lithology
T 1	17,00	7,70	1526,0	12,82	88,7	685,0	125,0	0,5	bas. komatiite
T 2	31,67	3,62	2503,0	13,01	112,0	2196,0	488,0	0,4	perid. komatiite
T 3	8,46	11,52	86,9	12,53	66,0	67,2	n.d.	0,9	tholeiite
T 3/1	9,17	11,30	173,0	12,75	52,6	66,6	n.d.	2,1	tholeiite
T 3/2	8,88	11,01	129,0	12,30	58,7	56,3	n.d.	1,4	tholeiite
T 4	35,07	2,89	3693,0	12,69	116,0	2077,0	469,0	1,5	perid. komatiite
T 5	12,39	11,22	602,0	12,46	58,6	196,0	61,3	1,4	bas. komatiite
T 6	39,55	2,61	1269,0	12,12	96,3	2102,0	306,0	0,6	perid. komatiite
T 7	7,67	13,70	105,0	15,34	58,2	67,4	14,3	0,9	tholeiite
T 8/1	32,46	3,05	1687,0	12,12	97,7	1802,0	343,0	0,6	perid. komatiite
T 9	13,94	6,12	1206,0	12,32	64,2	174,0	n.d.	0,9	bas. komatiite
T 10	33,58	3,18	3426,0	13,27	113,0	1697,0	313,0	0,9	perid. komatiite

+ total Fe as Fe_2O_3

method employed was modified after the procedure described by Seinnes (1971).

Great care was taken to prevent sample contamination. The whole rock samples were crushed between highly pure alumina plates, and from the crushed pieces the freshest portions were selected and pulverized in an agate mortar used for this purpose only. The pulverized samples were carefully homogenized before they were weighed into alumina capsules for irradiation.

Irradiation lasted three days and was carried out in the nuclear reactor FRJ-1 of the Kernforschungsanstalt Jülich. After six days cooling time the activated samples were transferred into plastic capsules for measurement of their gamma-ray spectra. Measurements were carried out three times within a time period of three months using a 70 cm^3 Ge(Li) detector. For increased sensitivity an anti-coincidence-shielded Ge (Li) detector was used for samples containing Au values below the detection limit of the 70 cm^3 Ge(Li) detector which was 2,0 ppb Au.

To test the IENAA method, 30 samples were also analysed for their Au content by a radiochemical, carrier-based neutron activation technique (RNAA). The radiochemical procedure used was a hybrid scheme which closely follows that described by Crocket *et al.* (1968).

To estimate the reproducibility of the Au analyses by the two NAA methods employed, sample HP 124 was separately analysed three times by RNAA and seven times by IENAA (Table 6). The precision obtained was better than 8 per cent for the IENAA method, and better than 17 per cent for the RNAA method.

To assess the accuracy of the gold determinations, three samples of the USGS rock standard PCC-1 were analysed by RNAA. The following Au contents were obtained: 1,4 ppb, 1,3 ppb, and 1,4 ppb. Flanagan (1973) reported a maximum Au content of 1,6 ppb for the PCC-1 standard.

IV. RESULTS

A. Data from the Archaean Mafic and Ultramafic Rocks of Southern Africa

1. Gold distribution

The distribution of the Au concentrations found in mafic and ultramafic rock types from southern Africa are shown as histograms in Fig. 4. The raw data are positively skewed and it is apparent that the gold values generally exhibit quasi log-normal distributions. The 64 *mafic* rocks possess an arithmetic mean ($\bar{x}$) Au content of 6,9 ppb with a range from 0,1 to 71,7 ppb Au and a geometric mean (G) Au content of 1,8 ppb. The 34 *ultramafic* rocks vary from 0,3 to 372 ppb Au, with $\bar{x}$ = 17,9 ppb Au, and G = 2,5 ppb Au. If the extreme value of 372 ppb Au, found in a serpentinized peridotitic "unsettled spinifex rock" from the Belingwe Belt (sample NG 138), is removed from the data set, $\bar{x}$ for the ultramafic rocks changes to 7,2 ppb Au, and G to 2,2 ppb Au.

A non-parametric Kolmogorov-Smirnov two-sample test (Miller and Kahn, 1965; Siegel, 1956) of Au values for mafic and ultramafic rocks shows that the d-value (0,09) does not exceed the critical value of D (0,259) at a 0,10 level of significance). Thus, the hypothesis that the pair of samples investigated has been drawn from the same population is accepted. However, if the samples are grouped according to their provenance areas (Rhodesian Craton *versus* Kaapvaal Craton) the d-value (0,288) exceeds the critical value of D (0,274) at a 0,05 level of significance. Consequently the hypothesis that the samples have been drawn from the same population is rejected.

The cumulative frequency distribution of the gold data from mafic and ultramafic rocks plotted on logarithmic probability paper does not reveal a straight line but shows a definite curvature with an inflection point, thereby indicating the presence of two overlapping populations in the data set (Fig. 5). These two populations can be termed a "background value population" and an "excess value population" (Lepeltier, 1969). Applying the graphical partitioning method for bimodal probability curves, as described by Sinclair (1978), it was found that 19 per cent of the gold data belongs to an "excess population B ($\bar{x}_B$= 33 ppb Au ; $\bar{x}_B+2s_L$=155 ppb Au; $\bar{x}_B-2s_L$=6,8 ppb Au), and 81 per cent to a "background population" A ($\bar{x}_A-2s_L$=0,1 ppb Au) (Fig. 5).

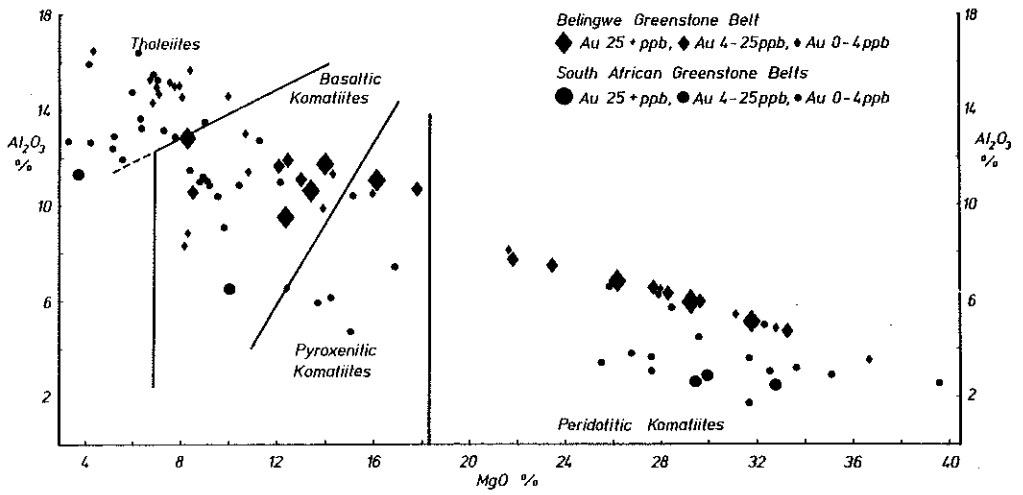

Figure 3 : *Al_2O_3 vs MgO diagram (Viljoen and Viljoen, 1979) of data from greenstone terranes. The diagram clearly reveals the higher Au concentrations of the samples from the Belingwe Belt and the Al_2O_3 depletion of the samples from the South African greenstone belts.*

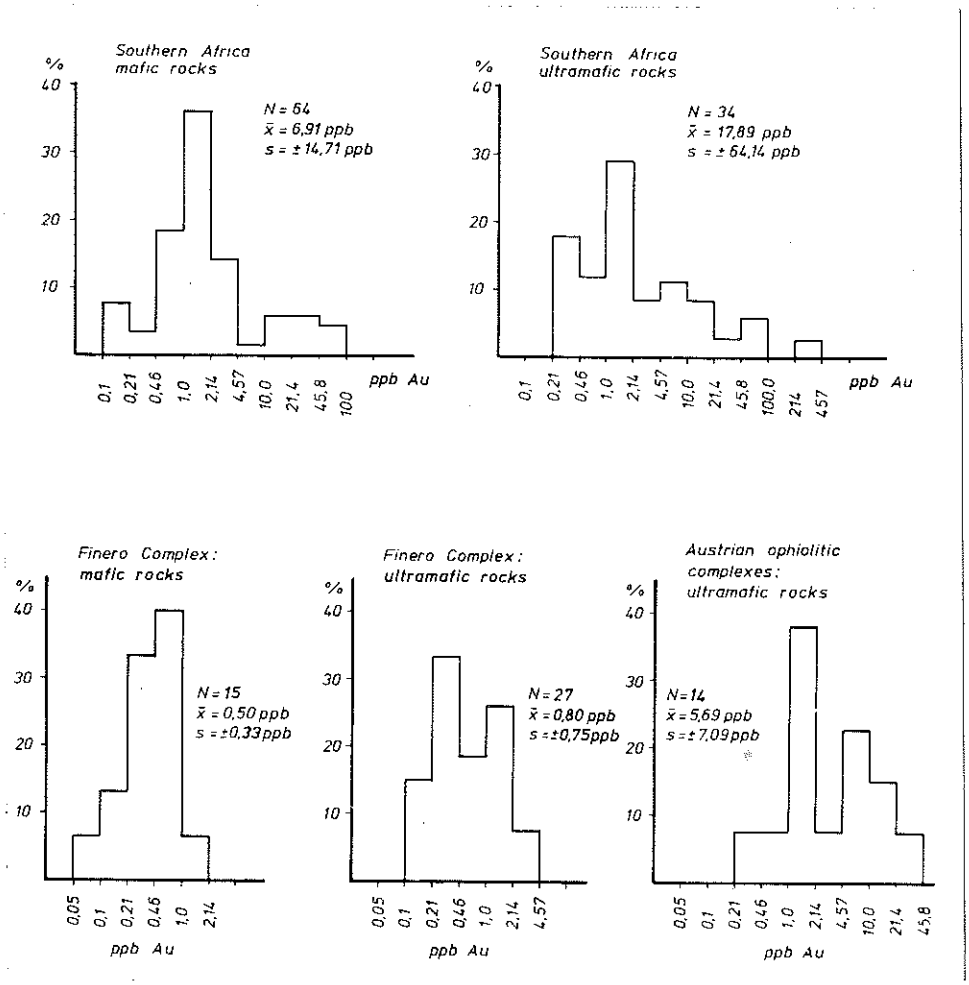

Figure 4 : *Distributions of Au/values in the various areas sampled in this study.*

TABLE 2

GEOCHEMICAL DATA FOR SAMPLES COLLECTED IN THE EERSTELING AREA OF THE PIETERSBURG GREENSTONE BELT

Sample	MgO %	Al_2O_3 %	Cr ppm	Fe_2O_3+ %	Co ppm	$Ni_{tot.}$ ppm	$Ni_{sulf.}$ ppm	Au ppb	Lithology
PM 16	15,54	4,90	1239,0	11,92	76,8	318,0	42,5	2,5	serpentinite
PM 31	26,64	3,77	3282,0	18,48	137,0	1141,0	406,0	1,3	serpentinite
PM 34	4,17	15,98	275,0	3,27	70,2	275,0	22,5	2,0	quartz porphyry
PM 42	5,22	12,59	252,0	15,51	64,3	114,0	30,0	2,9	quartz porphyry
PM 83	3,81	11,52	168,0	18,57	62,0	73,6	34,0	16,8	basalt (pyrite rich)
PM 132	15,82	10,84	1547,0	11,44	60,0	668,0	38,8	0,8	basalt
PM 133	31,58	14,50	4626,0	13,52	145,0	2436,0	8,6	1,5	serpentinite
PM 142	3,37	12,93	160,0	11,35	32,6	41,8	n.d.	2,4	basalt
PM 145	5,27	13,05	108,0	12,72	55,5	71,0	n.d.	3,5	basalt
M 112	5,62	12,12	354,0	11,08	59,4	123,0	21,3	1,8	schistose basalt
M 115	6,00	14,68	100,0	13,55	42,5	61,8	n.d.	2,3	schistose basalt
77/53	14,37	6,06	1359,0	12,63	53,2	195,0	11,3	0,4	basalt
77/59	28,43	5,69	2335,0	11,08	88,5	1658,0	71,3	0,3	komatiite
77/61	29,63	4,47	2054,0	9,09	84,0	1754,0	218,0	0,3	komatiite
77/62	32,31	4,94	1734,0	9,06	80,7	1454,0	394,0	0,3	komatiite
77/63	25,90	6,65	2093,0	9,77	81,5	1571,0	155,0	0,3	komatiite
77/64	9,91	9,21	1273,0	10,92	68,7	227,0	n.d.	1,0	komatiite
77/79	9,02	11,28	927,0	11,13	55,9	225,0	10,0	0,8	komatiite
77/80	10,56	11,03	541,0	12,85	64,9	235,0	10,0	0,9	komatiite
77/81	10,25	6,69	1601,0	13,26	67,2	204,0	17,5	20,2	komatiite
228768	25,40	3,46	2844,0	12,41	76,8	952,0	350,0	1,5	pyroxenite sill
228769	29.97	2,74	2972,0	13,53	99,0	1036,0	270,0	5,3	pyroxenite sill
2287611	30,45	2,98	2980,0	14,38	104,0	1215,0	375,0	8,1	pyroxenite sill
2287612	26,65	3,01	2939,0	17,75	77,6	986,0	669,0	3,8	pyroxenite sill
2287614	7,27	16,42	382,0	11,84	68,9	122,0	21,3	1,0	basalt
2287615	27,57	3,68	2791,0	17,24	66,2	816,0	375,0	1,4	pyroxenite sill
2287616	4,30	12,99	199,0	12,57	66,8	71,2	n.d.	1,0	basalt
2287618	32,38	2,59	4471,0	13,24	179,0	3092,0	613,0	4,1	serpentinite
258762A	11,64	13,11	752,0	10,32	52,4	187,0	16,3	0,4	komatiite
258762B	9,12	13,70	581,0	9,33	46,3	158,0	16,3	0,6	komatiite
268761	9,54	10,43	954,0	12,18	75,9	239,0	16,3	2,0	basalt
268762	6,65	13,80	207,0	16,97	55,4	92,9	n.d.	0,5	basalt
268763	6,31	13,60	221,0	13,07	61,8	213,0	16,3	1,8	pillow lava
278768	8,04	13,14	461,0	9,09	63,5	314,0	16,3	3,4	tholeiite
278769	7,07	15,54	369,0	12,42	53,1	140,0	33,8	1,1	schistose basalt

+ total Fe as Fe_2O_3

If the threshold value for the "excess population" is set at $\bar{x}_B$-2s=6,8 ppb Au, 3 values from South African greenstone belts and 17 values from the Belingwe greenstone belt make up the "excess population". The "excess population" is thus, almost exclusively determined by the Belingwe Belt samples (85%), whereas 44 values (56%) from South African greenstone belts, and 34 values (44%) from the Belingwe greenstone belt contribute to the "background population". This underlines the result of the Kolmogorov-Smirnov statistics and demonstrates the presence of a regional difference in the gold content of ultramafic and mafic volcanic rocks between the Belingwe Belt of Zimbabwe and the greenstone belts of South Africa — the volcanics of the former possessing higher levels of gold concentration.

Furthermore, all samples for which major element data were available have been given different symbols according to their origin (South African greenstone belts and Belingwe greenstone belt) and the magnitude of their gold concentrations. The data were then plotted in the Al_2O_3 *versus* MgO diagram (Fig. 3) suggested by Viljoen and Viljoen (1979). This representation also confirms the generally higher gold concentrations of the Belingwe samples and, in addition, it reveals distinctly lower Al contents for the samples originating from the South African greenstone belts.

2. *Correlation of Gold with Major and Trace Elements*

To study correlations of gold with other elements, the samples from all the greenstone terranes investigated have been treated as one population. For the calculation of the product moment correlation coefficients (Table 7), and subsequent statistical tests, the Cr, $Ni_{tot.}$, Co, and Au data have been logarithmically transformed. For the major elements the raw data were used.

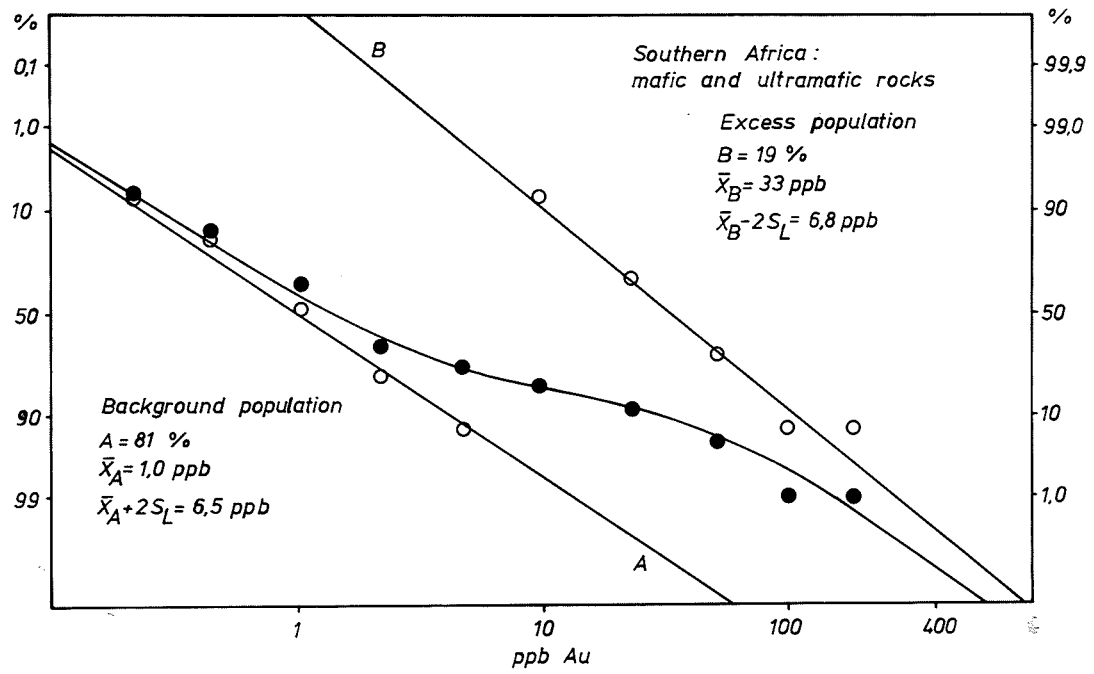

Figure 5 : *Partitioned log probability plot of 98 Au-values for volcanics from southern African greenstone terranes. $\bar{x}_A$= mean value of "background population"; $\bar{x}_B$-2s_L= threshold value for anomalous gold concentrations. For further explanations see text.*

The correlation coefficients for the pairs Au-MgO, Au-Fe, Au-Cr, Au-Co, and Au-$Ni_{tot.}$ were not significantly different from zero at the 5 per cent level. Absence of correlation between gold and major elements is furthermore demonstrated by the scattergram of Au *versus* the modified Larsen factor (1/3Si+K-Ca-Mg) (see Nockolds and Allen, 1953) which does not indicate a trend for the Au values (Fig. 6). This scattergram also supports the threshold level of 6,8 ppb Au (Fig. 5) defining a "base line" below which about 80 per cent of the gold data are concentrated.

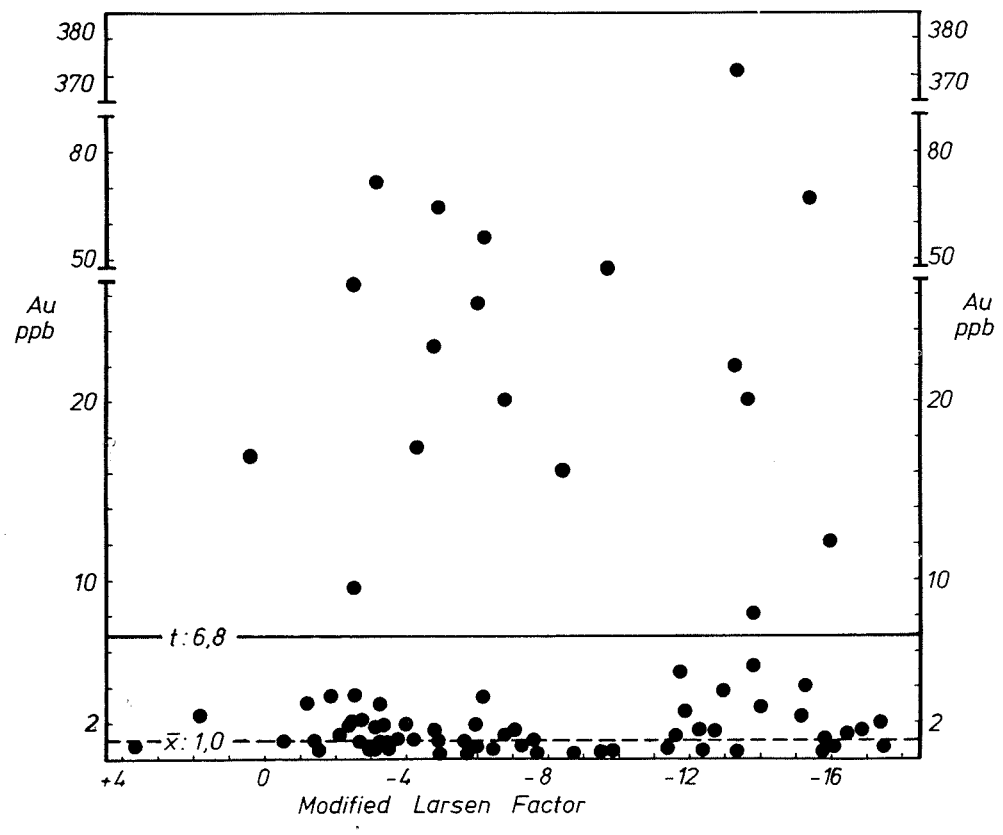

Figure 6 : *Plot of modified Larsen factor versus Au concentrations of 98 samples from southern African greenstone belts. The threshold value t=6,8 ppb Au, and the mean value $\bar{x}$=1,0 ppb Au of the "background population" are the same as in Fig. 5.*

This pattern of gold data corresponds with the findings of Fritze and Robinson (1969), Stephenson and Ehmann (1971), Tilling *et al.* (1972), and Keays and Scott (1976), who also observed in their data sets that numerous samples possess low Au-values and only a few samples have "excess values". This, of course, is in agreement with the observations of Ahrens (1957), who demonstrated that most trace elements follow a log-normal

TABLE 3

GEOCHEMICAL DATA FOR SAMPLES COLLECTED IN THE NGESI RIVER AREA OF THE BELINGWE GREENSTONE BELT

Sample	MgO %	Al_2O_3%	Cr ppm	Fe_2O_3+%	Co ppm	$Ni_{tot.}$ ppm	$Ni_{sulf.}$ ppm	Au ppb	Lithology
AT 4	n.d.	n.d.	2870,0		110,0	1771,0	285,0	2,1	perid. komatiite
AT 6	10,79	11,45	897,0	7,83	48,1	163,0	21,3	1,1	spinifex rock
AT 7	21,60	8,12	2587,0	9,02	86,3	955,0	146,0	0,4	perid. komatiite
AT 8	31,13	5,49	2926,0	9,40	114,0	2053,0	261,0	2,3	ultramaf.ridge
AT 9	36,69	3,48	4195,0	11,52	135,0	2116,0	875,0	1,8	ultramaf.ridge
AT 11	12,97	11,14	1516,0	9,18	63,9	410,0	241,0	23,2	bas.komatiite
u/c	n.d.	n.d.	792,0	n.d.	54,1	149,0	10,0	1,1	bas.komatiite
B 4	27,93	5,30	2700,0	10,45	109,0	1718,0	12,5	1,8	spinifex flow
BL 1	7,57	15,25	443,0	6,41	45,0	163,0	33,8	1,6	massive lava
BL 6	n.d.	n.d.	947,0	n.d.	62,0	357,0	n.d.	0,8	massive lava
BL 8	8,32	8,84	958,0	12,37	63,0	378,0	21,3	0,1	massive lava
BL 9	12,39	6,63	1022,0	13,33	75,2	377,0	40,0	0,1	spherul.pillow
BL 14	n.d.	n.d.	440,0	n.d.	66,1	174,0	40,0	2,0	pillow lava
BL 18	6,68	15,27	202,0	7,96	36,6	92,6	28,3	3,1	pillow lava
BL 19	4,34	16,51	130,0	12,43	37,2	28,7	17,5	0,7	spherul.pillow
BL 22	8,38	15,72	30,9	10,56	48,9	172,0	40,0	0,6	massive flow
BL 25	n.d.	n.d.	413,0	n.d.	47,0	145,0	46,3	7,4	pillow lava
BL 26	7,09	14,71	327,0	13,62	57,2	145,0	40,0	1,0	pillow lava
BL 27	7,03	14,99	323,0	12,63	63,4	158,0	62,5	1,0	pillow margin
BL 28	7,92	14,99	336,0	12,66	65,4	159,0	40,0	0,5	pillow interior
BL 29	8,15	8,35	593,0	9,98	73,4	112,0	17,5	0,1	lava flow
BL 31	6,85	14,36	155,0	10,13	54,2	57,9	20,0	0,5	lava flow
BL 32	14,37	11,35	1090,0	6,89	46,7	302,0	27,5	0,1	sill
BL 33	10,04	14,64	580,0	8,88	63,1	162,0	21,3	0,5	spherul.pillow
BL 34	10,74	13,06	605,0	8,41	68,9	170,0	21,3	1,2	pillow lava
BL 35	n.d.	n.d.	513,0	n.d.	69,7	74,1	21,3	1,1	pillow lava
NG 4	13,94	9,88	994,0	13,36	67,0	324,0	16,3	1,8	column.spinifex
NG 12	14,17	11,24	1405,0	8,16	61,1	422,0	67,5	26,5	pillow lava
NG 63	12,08	11,73	1182,0	11,44	66,1	335,0	n.d.	17,6	graded tuff
NG 110	7,76	15,09	390,0	12,52	44,3	84,1	30,0	0,6	massive flow
NG 120	17,80	10,77	706,0	13,61	97,0	233,0	32,5	16,2	pillow lava
NG 121	16,06	10,54	1865,0	11,93	79,3	687,0	78,8	1,5	pillow lava
NG 124	32,84	4,88	2658,0	8,47	103,0	1843,0	164,0	0,6	ultramaf.ridge
NG 133a	11,50	11,12	1226,0	11,44	65,0	183,0	n.d.	3,1	spinifex flow
NG 137	13,28	10,65	1760,0	12,37	66,6	213,0	41,3	25,4	pillow lava
NG 138	26,09	6,80	2123,0	12,27	107,0	1488,0	141,0	372,0	ultramaf.ridge
NG 152	31,81	5,08	2986,0	9,55	112,0	1933,0	356,0	66,9	ultramaf.ridge
NG 157	28,18	6,34	2714,0	10,61	102,0	1516,0	419,0	20,2	ultramaf.flow
NG 159	8,16	12,86	717,0	11,18	48,8	159,0	46,3	71,7	crosscutting dyke
NG 161	8,09	14,58	444,0	12,03	48,5	102,0	57,2	3,5	lava flow
NG 162	12,37	9,64	1292,0	10,47	66,2	457,0	104,0	64,3	cleaved mafic
NG 172	n.d.	n.d.	184,0	n.d.	27,1	127,0	46,3	29,6	massive flow
NG 198	8.46	10,66	1033,0	9,37	59,0	228,0	128,0	9,5	massive flow
NG 199	21,71	7,77	2327,0	12,55	79,1	953,0	62,5	4,8	massive flow
NG 207a	27,07	6,43	2618,0	11,79	93,6	1358,0	n.d.	0,6	ultramaf.flow
NG 208	29,57	6,06	2401,0	11,53	96,7	1440,0	n.d.	13,2	ultramaf.flow
NG 212	27,80	6,50	2629,0	11,16	100,0	1467,0	625,0	22,4	ultramaf.flow
NG 214	23,25	6,74	2594,0	12,66	79,9	1107,0	n.d.	5,6	ultramaf.pillow
NG 217	29,34	6,04	2365,0	11,69	92,3	1441,0	n.d.	48,2	ultramaf.flow
NG 219	33,20	4,73	2928,0	7,23	103,0	1723,0	375,0	12,2	ultramaf.ridge
NG 220	16,15	11,12	1990,0	11,99	70,6	628,0	144,0	56,2	spinifex flow

+ total Fe as Fe_2O_3

Major element data have been obtained from Nisbet *et al.* (1977)

TABLE 4

GEOCHEMICAL DATA FOR THE FINERO ULTRAMAFIC COMPLEX

Sample	MgO %	Al_2O_3 %	S %	Cr ppm	$Fe_2O_3^+$ %	Co ppm	Ni ppm	Cu ppm	Au ppb	Lithology
UA 33	10,94	18,71	0,047	457,0	9,49	46,7	83,2	28,0	0,3	amphibolite
UA 80	10,93	19,97	0,014	866,0	5,21	33,3	95,5	17,0	0,3	amphibolite
UA 84	12,45	21,89	n.d.	378,0	6,02	52,3	204,0	21,0	0,9	amphibolite
UA 92	17,50	17,20	0,041	827,0	4,68	52,9	358,0	52,0	1,3	amphibolite
UA 115	8,73	21,70	0,024	885,0	7,69	51,4	151,0	60,0	0,7	amphibolite
UA 150	n.d.	n.d.	n.d.	243,0	8,14	46,3	62,6	n.d.	0,1	amphibolite
UA 155	n.d.	n.d.	0,068	478,0	6,02	46,3	135,0	n.d.	0,6	amphibolite
UA 160	n.d.	n.d.	0,062	361,0	3,92	29,5	61,5	n.d.	0,4	amphibolite
OA 42	n.d.	n.d.	0,049	239,0	12,39	59,8	114,0	n.d.	0,7	amphibolite
OA 46	n.d.	n.d.	0,007	700,0	5,84	63,2	104,0	n.d.	0,6	amphibolite
OA 72	8,01	11,97	0,056	481,0	12,08	53,5	187,0	50,0	0,8	amphibolite
OA 74	11,99	13,50	0,018	210,0	13,49	43,3	35,5	32,0	0,3	amphibolite
OA 109	8,87	15,22	0,020	324,0	15,24	44,9	56,8	34,0	0,3	amphibolite
OA 133	n.d.	n.d.	0,007	211,0	12,06	45,4	36,8	n.d.	0,2	amphibolite
OA 145	n.d.	n.d.	0,031	361,0	5,58	54,4	71,2	n.d.	0,2	amphibolite
HP A	n.d.	n.d.	0,860	3533,0	10,12	230,0	11521,0	670,0	3,0	serpentinite
HP 4	44,80	0,70	0,008	14422,0	8,90	216,0	3685,0	17,0	1,0	peridotite
HP 21/1	n.d.	n.d.	0,340	4247,0	13,70	230,0	2896,0	225,0	1,2	serpentinite
HP 35	n.d.	n.d.	0,550	3441,0	13,45	182,0	2214,0	310,0	1,1	serpentinite
HP 36	n.d.	n.d.	0,610	1889,0	10,07	141,0	2074,0	385,0	1,3	peridotite
HP 39	40,95	2,00	0,024	3425,0	10,05	113,0	2075,0	55,0	0,3	peridotite
HP 53	n.d.	n.d.	0,041	3420,0	8,23	106,0	2058,0	n.d.	0,4	serpentinite
HP 64	n.d.	n.d.	1,050	3685,0	6,95	139,0	4793,0	390,0	1,6	serpentinite
HP 79	43,12	1,75	0,280	6401,0	12,63	173,0	2287,0	220,0	1,0	peridotite
HP 81	40,78	3,06	0,026	3603,0	11,35	125,0	1246,0	90,0	0,3	peridotite
HP 93	44,09	1,18	0,015	3509,0	8,21	112,0	1920,0	25,0	0,4	peridotite
HP 96	n.d.	n.d.	0,250	6341,0	10,50	164,0	2632,0	360,0	1,2	peridotite
HP 101	45,11	2,10	n.d.	2974,0	10,14	184,0	2397,0	22,0	0,5	peridotite
HP 118	38,35	2,25	0,130	2958,0	6,24	128,0	1727,0	15,0	0,3	serpentinite
HP 122	n.d.	n.d.	0,100	3157,0	6,45	103,0	3133,0	n.d.	1,2	serpentinite
HP 124	n.d.	n.d.	0,910	2400,0	7,02	212,0	7250,0	890,0	2,8	serpentinite
HP 146	n.d.	n.d.	0,040	2425,0	6,17	115,0	1933,0	n.d.	0,2	serpentinite
HP 153	n.d.	n.d.	0,260	4247,0	7,39	156,0	3141,0	n.d.	1,4	peridotite
PP A	n.d.	n.d.	0,011	2218,0	7,91	115,0	1935,0	n.d.	0,2	peridotite
PP 27	44,50	0,74	0,003	2662,0	8,19	110,0	1927,0	10,0	0,4	peridotite
PP 30	n.d.	n.d.	0,003	2692,0	7,99	123,0	2273,0	n.d.	0,2	peridotite
PP 87	47,14	0,34	0,005	2506,0	8,24	163,0	2226,0	10,0	0,5	peridotite
PP 88	43,72	0,71	0,004	3039,0	7,39	109,0	1925,0	5,0	0,3	peridotite
PP 89	n.d.	n.d.	0,004	2772,0	7,04	107,0	2121,0	n.d.	0,2	peridotite
PP 144	n.d.	n.d.	0,004	2220,0	7,84	119,0	2097,0	n.d.	0,1	peridotite
PP 152	n.d.	n.d.	n.d.	2772,0	7,99	119,0	2193,0	n.d.	0,5	peridotite
PP 158	n.d.	n.d.	0,005	2246,0	8,81	135,0	2019,0	n.d.	0,2	peridotite

$^+$ total Fe as Fe_2O_3

distribution. Fritze and Robinson (1969) postulated a "two component model" for the Au distribution pattern and proposed that Au occurs in two components, viz. in (an) abundant mineral phase(s) of low Au contents responsible for the "background population", and (an) erratically distributed phase(s) of higher Au contents contributing the "excess population".

However, statistically significant positive correlation coefficients were observed for the pairs Ni-MgO, Ni-Co, Ni-Cr, and Cr-MgO (Table 7). These positive correlations are also manifested by the distinct increase of Cr and Ni contents with decreasing values of the modified Larsen factor (Fig. 7a, b). This indicates that in the samples investigated from southern African Ni, (and probably less pronounced Cr) is predominantly contained in silicate minerals such as olivine, pyroxene, hornblende, and micas. Chromite has been found to be a relatively abundant constitutent in the peridotitic samples, whereas Ni-sulphides were only rarely observed. The scarcity of Ni-sulphides

TABLE 5

GEOCHEMICAL DATA FOR THE HOCHGRÖSSEN (S SAMPLES) AND KRAUBATH (K SAMPLES) ULTRAMAFIC COMPLEXES

Sample	MgO %	Al_2O_3 %	Cr ppm	$Fe_2O_3^+$ %	Co ppm	$Ni_{tot.}$ ppm	Au ppb	Lithology
K 5 A	41,50	1,81	1765,0	12,10	121,0	1097,0	1,0	serpentinite
K 8 C	41,50	0,48	3650,0	11,10	118,0	1676,0	1,7	serpentinite
K 15	37,35	1,99	5051,0	12,11	173,0	3408,0	25,0	serpentinite
K 35	31,50	0,43	4480,0	17,72	162,0	2013,0	12,6	serpentinite
S 9	n.d.	n.d.	272,0	n.d.	58,4	95,1	4,6	amphibolite
S 42	n.d.	n.d.	3967,0	n.d.	160,0	1817,0	1,5	talc-carbonate
S 45	n.d.	n.d.	618,0	n.d.	38,4	1979,0	0,4	talc-carbonate
S 76	42,33	0,19	2621,0	9,10	126,0	1976,0	1,1	serpentinite
S 96	n.d.	n.d.	253,0	n.d.	53,2	70,4	8,6	amphibolite
S 100	41,50	0,71	3277,0	10,61	110,0	2163,0	2,0	serpentinite
S 116	37,52	1,99	6963,0	14,20	154,0	2477,0	13,8	serpent.dunite
S 118	41,50	1,57	9072,0	14,81	157,0	3090,0	4,6	serpent.dunite
S 122	35,28	1,33	5359,0	12,90	110,0	1894,0	2,2	serpent.mylonite
S 169	n.d.	n.d.	47,6	n.d.	35,2	493,0	0,6	steatite

$^+$ total Fe as Fe_2O_3

is furthermore underlined by the low ratios of $Ni_{sulf.}/Ni_{tot.}$ obtained for most of the samples, with the serpentinized samples usually possessing the highest $Ni_{sulf.}/Ni_{tot.}$ ratio.

3. Factor Analysis

To further investigate the interrelation of the trace element data of the mafic and ultramafic rocks, an R-mode factor analysis of Mg, Cr, Fe, Co, $Ni_{tot.}$, and Au data was undertaken. The product moment correlation coefficient matrix analysed by the R-mode factor analysis is given in Table 7, and the varimax rotated factor matrix together with the communality values is presented in Table 8.

The result of the analysis reveals that a three factor model accounts for 92,2 per cent of the total variance in the original data set. The high communality values displayed by all investigated elements show that each of them is explained by the three factor model to more than 85 per cent, with Cr and Co having the lowest communalities.

TABLE 6

COMPARISON OF GOLD DETERMINATIONS BY THE IENAA AND RNAA METHODS

IENAA ppb Au	RNAA ppb Au
3,2	3,3
3,1	2,5
2,9	2,5
3,1	
2,6	
2,7	
2,8	

(SAMPLE HP 124)

Factor 1 possesses high loadings of Mg, Cr, $Ni_{tot.}$, and Co. As these elements are typically associated with ultramafic rocks, factor 1 is named "ultramafic factor". *Factor 2* is almost exclusively made up of Fe and, thus, is termed "Fe factor". Iron probably forms a separate factor from factor 1 because the Fe contents of the investigated mafic ($\bar{x}$= 11,46%) and ultramafic rocks ($\bar{x}$= 11,47%) are virtually identical, as opposed to

TABLE 7

CORRELATION MATRIX FOR THE GEOCHEMICAL DATA FROM SAMPLES COLLECTED IN GREENSTONE TERRANES

	Mg	log. Cr	Fe	log. Co	log. $Ni_{tot.}$
log. Au	0,142	0,205	0,099	0,164	0,148
log. $Ni_{tot.}$	0,925	0,885	-0,082	0,862	
log. Co	0,829	0,758	0,097		
Fe	-0,021	-0,063			
log. Cr	0,829				

N=98; $r_{.99}$= 0,254

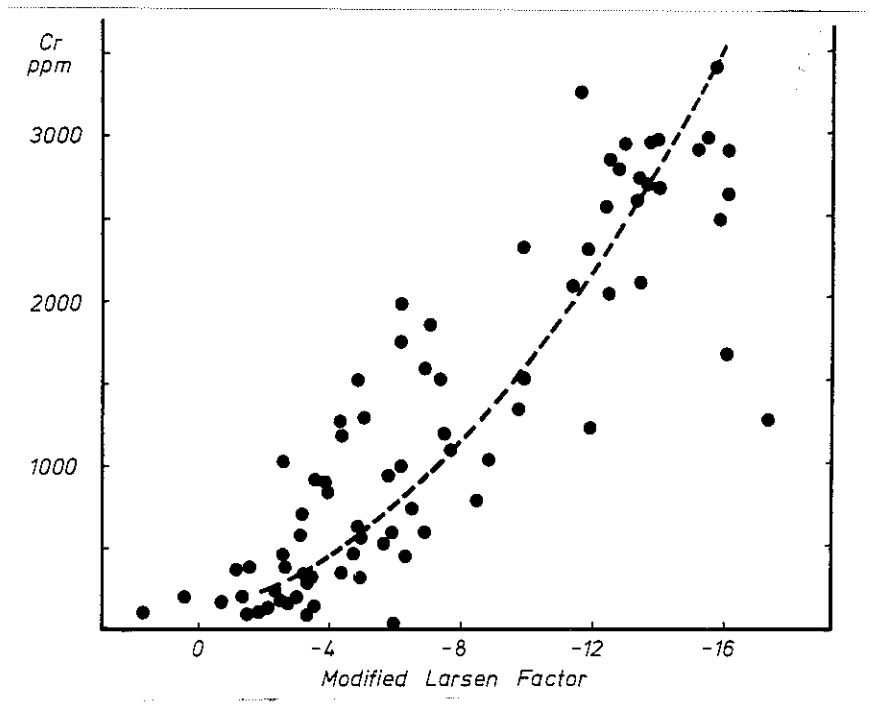

Figure 7a : Plot of modified Larsen factor versus Cr values of 98 samples from southern African greenstone belts. Note increase of Cr values with decreasing modified Larsen factor.

the mean Mg, Cr, Co, and Ni_{tot} contents which vary distinctly between the mafic and ultramafic samples. *Factor 3* is largely explained by Au and, therefore, termed "Au factor". The complete separation of Au from the other elements is interpreted to reflect the lack of any fundamental relationship between Au and any other variable investigated in the factor analysis.

B. Comparative Data from Palaeozoic European Ultramafic Complexes

1. Gold Distribution

The samples from the Finero Complex (27 peridotites and 15 amphibolites) possess a very limited range of Au concentrations (0,1 to 3,0 ppb). Gold in both peridotites and amphibolites is quasi log-normally distributed as indicated by the histograms of Fig. 4. The peridotites exhibit an arithmetric mean Au content of 0,8 ppb (G= 0,6 ppb Au) and the amphibolites an arithmetric mean Au content of 0,5 ppb (G= 0,4 ppb Au). A non-parametric Duncan "multiple range test" reveals that the hornblende peridotite (HP samples in Table 4) possesses a significantly (0,05 level) higher mean gold content than the phlogopite peridotite and the amphibolite. The latter two rock types do not differ significantly from each other in their mean gold contents. In this context it is interesting to note that the hornblende peridotite exceeds the other rock types of the Finero Complex in its greater content of sulphide minerals (pyrrhotite, pentlandite, chalcopyrite (Oberthür, 1978).

The range of the 14 samples from the Hochgrössen and Kraubath ophiolotic complexes varies from 0,4 to 25,0 ppb Au; their arithmetic mean content is 5,7 ppb Au (G= 2,9 ppb Au (Fig. 4).

The European complexes show Au distribution patterns similar to the greenstone belts from southern Africa in as much as they follow the earlier discussed "two component model" of Fritze and Robertson (1969).

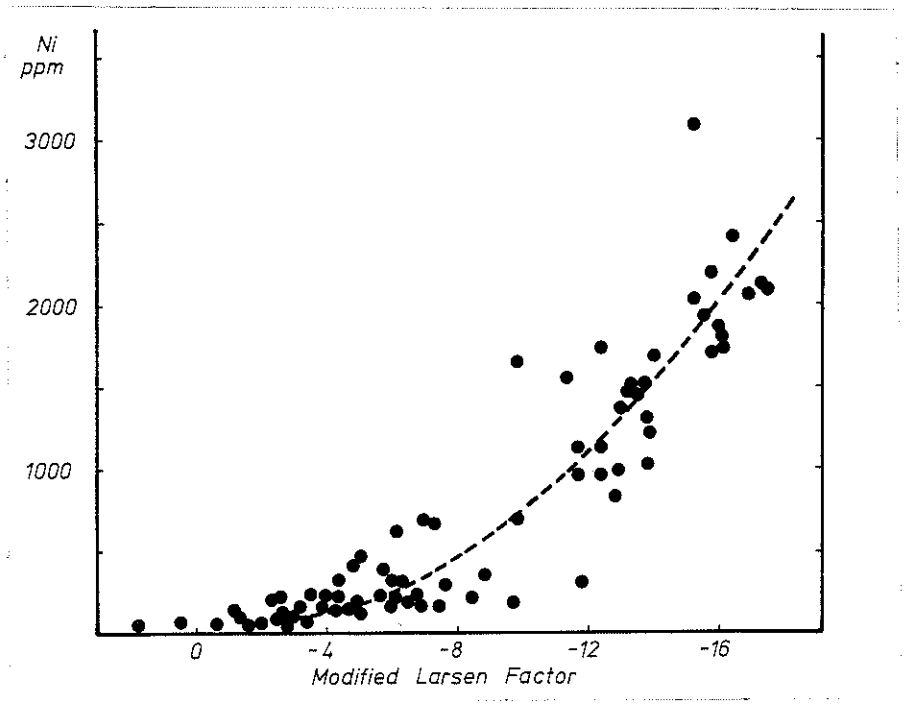

Figure 7b : Plot of modified Larsen factor versus $Ni_{tot.}$ values of 98 samples from southern African greenstone belts. Note increase of $Ni_{tot.}$ values with decreasing modified Larsen factor.

Threshold values or "base line levels" were determined separately for the ultramafic and mafic rocks from the Finero Complex and for the samples from the Austrian complexes using the partitioning method of Sinclair (1978). The "base line levels" are marked in the pertinent distribution plots of the Au values (Fig. 8). The peridotites from Finero show a slightly higher "base line level" (0,7 ppb Au than the associated amphibolite (0,6 ppb Au). All of the points lying above the "base line level" of the peridotite samples reflect values originating from the relatively sulphide-rich hornblende peridotite. The Kraubath and Hochgrössen values are rather inhomogeneous (Fig. 8), but a "base line level" (3,7 ppb Au) higher than the levels found for the Au values from the Finero Complex is indicated.

TABLE 8

R-MODE FACTOR MATRIX FOR THE GEOCHEMICAL DATA FROM SAMPLES COLLECTED IN GREENSTONE TERRANES

	Factor 1	Factor 2	Factor 3	Communality
Mg	0,95244	-0,01252	0,04222	0,90909
Cr	0,91244	-0,07751	0,13115	0,85575
Fe	-0,01370	0,99570	0,05089	0,99419
Co	0,91442	0,13502	0,06031	0,85804
$Ni_{tot.}$	0,97421	-0,07626	0,05351	0,95777
Au	0,10027	0,05186	0,99305	0,99889
Sum of squares	3,53518	1,o2432	1,01422	
Expl. variabil.	58,92	17,07	16,90	
Cum. expl. var.	58,92	75,99	92,89	

2. *Correlation of Gold with Trace Elements*

The geochemical data from the Finero Complex, in contrast to the data from southern Africa, show significant correlations (0,05 level) for the peridotites between the pairs Au-S, Au-Ni, Au-Co, and Au-Cu (Table 9a). For the amphibolites correlation (0,05 level) was found for the pair Au-Ni (Table 9b).

For the Finero samples, the significant positive correlations are largely determined by combinations of high Au values with above average concentrations of disseminated sulphide minerals. This behaviour is also reflected by Fig. 8 where the "excess gold populations" are made up by Au values originating from samples rich in sulphide minerals.

The samples from the Austrian ophiolitic complexes do not reveal significant correlation coefficients (Table 10). This is possibly a reflection of the inhomogeneity of the data set investigated.

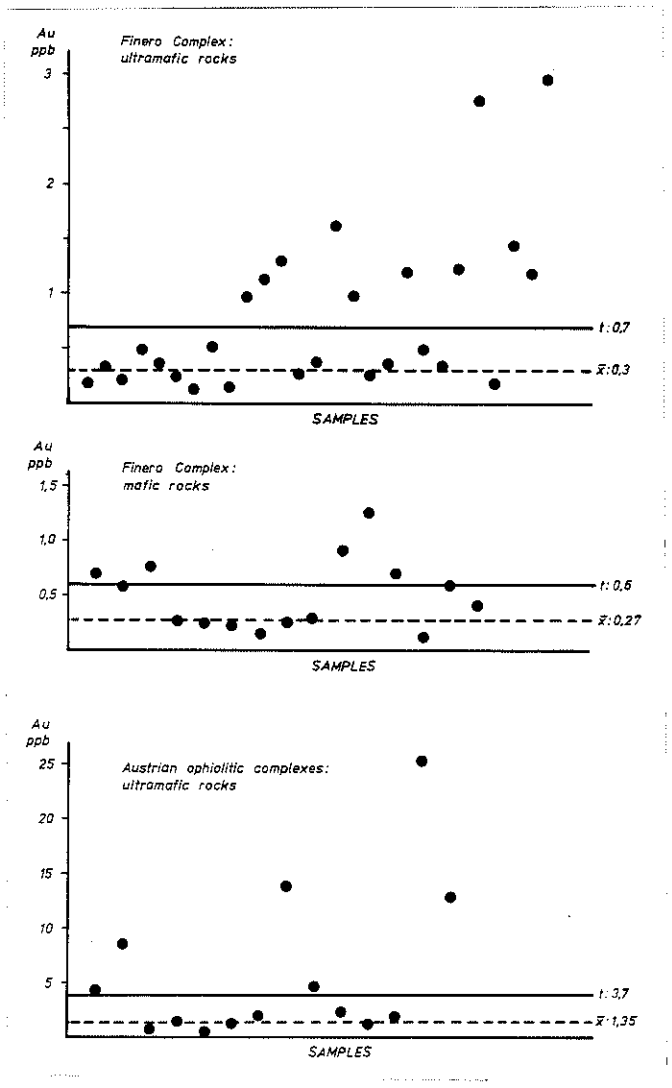

Figure 8 : Plot of Au concentrations of samples from ultramafic complexes of the European Alps. The threshold values (t) and the mean values of the "background populations" ($\bar{x}$) were obtained by partitioning of log probability plots.

C. Data from Early Precambrian Chemical Sediments of South Africa

The data (Cr, Co, Ni, Au, Th, U) of 32 samples of ferruginous cherts and oxide facies banded iron-formations are listed in Table 11. The Au data are also depicted in a histogram (Fig. 9). The Au-values have an extremely wide spread and range from 0,5 to 667 ppb with an arithmetic mean content ($\bar{x}$ of 129,9 ppb Au geometric mean G= 35,7 ppb Au). Gold particles were not detected optically and, thus, the mineralogical site of the gold is unknown. The gold values of the chemical sediments are distinctly higher than those of the investigated volcanic rocks. However, similar gold values ranging from 6 to 7836 ppb have been reported for detrital banded iron-formation and ferruginous chert fragments from the Kimberley Reef conglomerates of the Witwatersrand basin (Saager *et al.*, 1979).

Interflow sedimentary material from the komatiite sequence at Kambalda, Western Australia, studied by Bavinton and Keays (1978), have a range in Au values from 0,02 ppb to 1809 ppb Au ($\bar{x}$= 142,6 ppb Au). These chemical sediments are distinctly cherty in appearance and occur in a stratigraphic position similar to the material investigated from South Africa. In contrast to the banded iron-formations and ferruginous cherts from South Africa, the samples investigated from Western Australia are usually very rich in iron sulphides (20 to 25 wt%) with the non-sulphide fraction largely being made up of albite and quartz (Bavinton and Keays, 1978).

V. DISCUSSION

A. Gold Abundance

The results of this study demonstrate the presence of regional differences in the abundance of gold in mafic and ultramafic rock types. Such regional difference or specialization appears to camouflage any possible trends in the gold contents related to the differentiation attained by the rocks studied.

TABLE 9a

CORRELATION MATRIX FOR GEOCHEMICAL DATA FROM PERIDOTITE SAMPLES OF THE FINERO ULTRAMAFIC COMPLEX

	log.S	log.Cr	log.Co	log.Ni	log.Cu
log.Au	0,820	-0,057	0,693	0,746	0,822
log.Cu	0,920	-0,008	0,514	0,559	
log.Ni	0,549	0,211	0,630		
log.Co	0,558	0,450			
log.Cr	0,160				

for Au, Co, Ni, and Cr: N=27; $r_{.99}$= 0,470
for S: N=25; $r_{.99}$= 0,487
for Cu: N=17; $r_{.99}$= 0,575

TABLE 9b

CORRELATION MATRIX FOR GEOCHEMICAL DATA FROM AMPHIBOLITE SAMPLES OF THE FINERO ULTRAMAFIC COMPLEX

	log.S	log.Cr	log.Co	log.Ni	log.Cu
log.Au	0,348	0,507	0,357	0,820	0,511
log.Cu	0,536	0,212	0,497	0,345	
log.Ni	0,440	0,674	0,408		
log.Co	-0,005	0,048			
log.Cr	-0,359				

for Au, Co, Ni, and Cr: N=15; $r_{.99}$= 0,606
for S: N=13; $r_{.99}$= 0,641
for Cu: N=8; $r_{.99}$= 0,765

TABLE 10

CORRELATION MATRIX FOR GEOCHEMICAL DATA FROM THE HOCHGRÖSSEN AND KRAUBATH ULTRAMAFIC COMPLEXES

	log.Cr	log.Co	log.Ni
log.Au	0,379	0,501	-0,023
log.Ni	0,765	0,640	
log.Co	0,621		

N=15; $r_{.99}$= 0,606

Kolmogorov-Smirnov statistics (Miller and Kahn, 1968; Siegel, 1956) indicated, at a high level of significance (0,05), that the gold values of European and southern African samples were drawn from different populations (d=0,235; D=0,228), whereas the gold values of ultramafic and mafic rocks from southern Africa belong to a single population as shown earlier. Regional differences of gold abundance can also be demonstrated for the samples from southern Africa. If the gold data are grouped into a "Belingwe population" and "South African population", the Kolmogorov-Smirnov test indicates that a single population must be rejected (cf. Results). Even the gold data from the type locality of the Komati Formation in the Barberton Mountain Land and that of the Pietersburg greenstone belt do not show the same distribution patterns.

The regional differences in the abundance of gold may be attributed to:

(i) primary differences of gold abundance due to heterogeneous gold concentrations in the upper mantle, as suggested by Gottfried and Greenland (1972)

(ii) differences in crustal development and mantle differentiation, with primitive Archaean volcanics containing higher gold contents than equivalent younger rock types, as proposed by Viljoen *et al.* (1969) and

(iii) metamorphic and hydrothermal processes which affected the various areas sampled to different degrees and thereby changed and obscured primary gold distribution patterns as a result of secretion and redistribution processes (Boyle, 1961).

TABLE 11

TRACE ELEMENT CONTENTS IN FERRUGINOUS CHEMICAL SEDIMENTS FROM SOUTH AFRICA

Sample	Cr ppm	Ni ppm	Co ppm	Au ppb	Th ppm	U ppm	Group	Locality
77/2	439,0	7,78	16,8	150,0	n.d.	n.d.	Pietersburg	Rhenoster Koppies+
77/49	371,0	36,2	22,2	192,0	n.d.	0,08	Pietersburg	Amatava 41 KS+
77/50	466,0	26,6	13,9	151,0	n.d.	n.d.	Pietersburg	Amatava 41 KS+
77/51	390,0	189,0	23,2	3,7	n.d.	0,52	Pietersburg	Amatava 41 KS+
77/60A	397,0	46,6	19,7	173,0	n.d.	n.d.	Pietersburg	Landsberghoek 10 KS+
77/60B	566,0	64,3	16,2	170,0	n.d.	n.d.	Pietersburg	Landsberghoek 10 KS+
77/60C	356,0	47,1	12,8	157,0	n.d.	n.d.	Pietersburg	Landsberghoek 10 KS+
77/92	198,0	42,5	17,8	667,0	n.d.	1.04	Pietersburg	Eersteling 17 KS+
PM 112	n.d.	n.d.	n.d.	120,0	n.d.	n.d.	Pietersburg	Zandrivier 139+
PM 147	n.d.	n.d.	n.d.	80,0	n.d.	n.d.	Pietersburg	Turffontein 105+
228761	n.d.	n.d.	n.d.	145,0	n.d.	n.d.	Pietersburg	Rietvley 94+
228762	n.d.	n.d.	n.d.	202,0	n.d.	n.d.	Pietersburg	Rietvley 94+
218764	n.d.	n.d.	n.d.	88,0	n.d.	n.d.	Pietersburg	Zandrivier 139+
218767	n.d.	n.d.	n.d.	77,0	n.d.	n.d.	Pietersburg	Zandrivier 139+
218768	n.d.	n.d.	n.d.	495,0	n.d.	n.d.	Pietersburg	Zandrivier 139+
77/112	396,0	81,0	40,0	515,0	n.d.	n.d.	Onverwacht ?	Fumani Gold Mine++
77/112 11	n.d.	n.d.	n.d.	356,0	n.d.	n.d.	Onverwacht ?	Fumani Gold Mine++
77/117	448,0	17,7	8,0	45,4	n.d.	n.d.	Onverwacht ?	Fumani Gold Mine++
77/118	345,0	85,0	26,6	105,0	n.d.	0,33	Onverwacht ?	Fumani Gold Mine++
78/3	n.d.	n.d.	n.d.	24,6	n.d.	n.d.	Moodies	East of Fairview Mine+++
78/4	316,0	32,7	40,6	5,3	n.d.	n.d.	Fig Tree	East of Fairview Mine+++
78/9	259,0	158,0	81,1	40,5	n.d.	0,15	Onverwacht	Waaiheuvel 305+++
78/12	253,0	114,0	46,2	1,0	n.d.	1,07	Fig Tree	Schoonoord 25+++
78/20	332,0	13,4	38,2	0,5	1,82	0,39	Pongola	Redcliff 426 II
78-X	556,0	22,7	9,78	7,2	1,80	0,49	West Rand[1]	Evander Gold Field
78/23	522,0	10,2	3,10	5,4	0,27	0,07	West Rand[1]	Evander Gold Field
78/23 II	451,0	n.d.	19,5	0,6	1,75	0,64	West Rand[1]	Evander Gold Field
Bh-89	997,0	246,0	37,8	8,0	4,41	1,12	West Rand[1]	Frischgewaagd 142
Bh-206	559,0	31,5	4,80	162,0	0,93	0,38	West Rand[1]	Rietkuil 531
Bh-217 A	583,0	55,5	4,75	2,8	0,41	0,05	West Rand[1]	Langeverwacht 282
Bh-217 B	247,0	5,2	1,33	1,3	0,06	0,02	West Rand[1]	Langeverwacht 282
Bh-217 C	222,0	16,6	5,08	4,1	n.d.	0,03	West Rand[1]	Langeverwacht 282

+ Pietersburg Schist Belt ++ Sutherland Range +++ Barberton Mountain Land

[1] Contorted bed

Critical interpretation of the observed gold distributions, with respect to the above three hypotheses, allows postulation that the variations of the gold abundance are primary phenomena. These can best be explained by inhomogeneities of the gold content of the upper mantle and, are thus a reflection of the availability of gold during magma crystallization. The samples of volcanic rocks from the Belingwe and South African greenstone belts are, for example, chemically similar but the Belingwe samples have distinctly higher mean gold parameters than the South African samples (Table 12). Furthermore, the South African greenstone belts are some 500 Ma older than the Upper Greenstones of the Belingwe Belt (Jahn and Shih, 1974; Hawkesworth *et al.*, 1977) and consequently should have been derived from a less differentiated, and, according to Viljoen *et al.* (1969), a less gold depleted upper mantle.

To explain the differences in gold abundance as being due to metamorphic hydrothermal gold redistribution seems improbable. Samples from areas little altered by metamorphism (e.g. the Ngesi River area of the Belingwe Belt and the type locality of the Komati Formation in the Barberton Mountain Land) show marked differences in their gold abundance : the Belingwe samples having distinctly higher mean gold values than the material from the Barberton Mountain Land (Table 12). Anhaeusser *et al.* (1975) reported identical low gold contents for samples collected at the type locality of the Komati Formation (Table 12).

Numerous authors have attempted to establish the mean gold content of peridotitic and basaltic rocks on a worldwide scale (Turekian and Wedepohl, 1961; Vinogradov, 1962; Jones, 1969; Crocket, 1974; Boyle, 1979). The values quoted for ultramafic rocks range from 5 ppb Au (Vinogradov, 1962) to 11,4 ppb Au (Boyle, 1979) and for

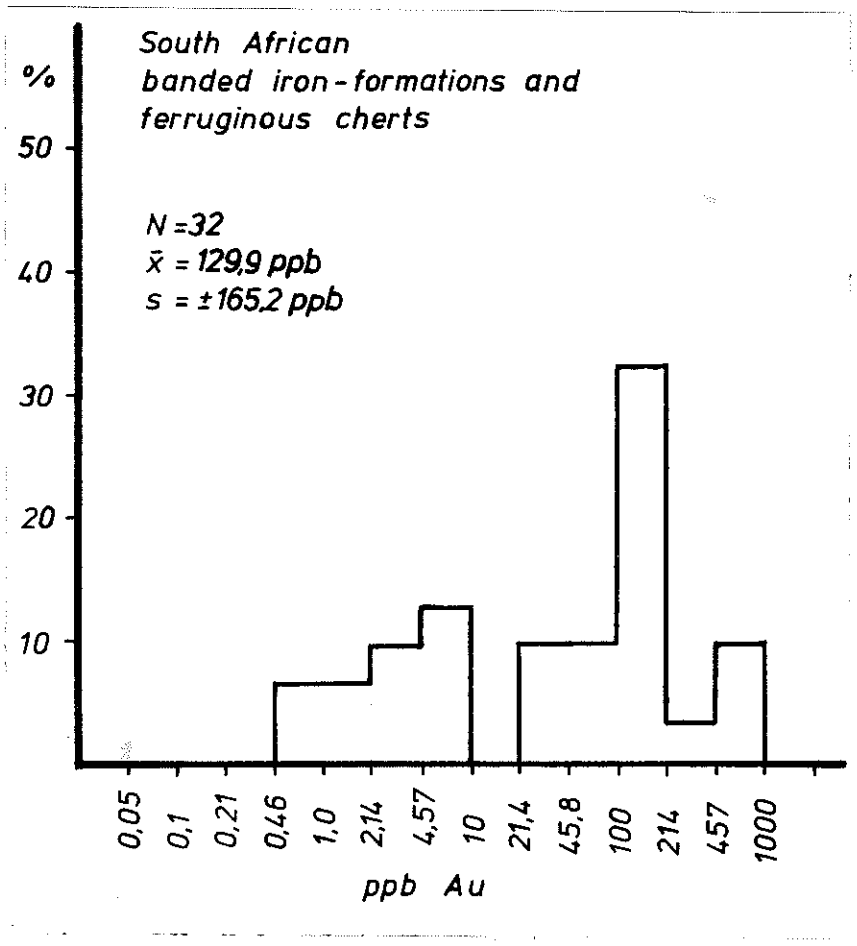

Figure 9 : Distribution of Au values in ferruginous chemical sediments from South Africa.

TABLE 12

ARITHMETIC MEAN ($\bar{x}$), STANDARD DEVIATION (s), GEOMETRIC MEAN (G), AND MEDIAN (Q) OF GOLD CONTENTS IN THE VARIOUS SAMPLE AREAS INVESTIGATED

Sample Area	Number of Samples	Range (ppb)	$\bar{x}$ (ppb)	s (ppb)	G (ppb)	Q (ppb)
Southern African greenstone belts : (Belingwe Belt, Pietersburg Belt, Barberton Mountain Land)	98	0,1-372	10,8	39,6	2,1	1,55
South African greenstone belts : (Pietersburg Belt, Barberton Mountain Land)	47	0,4-20,2	2,4	3,8	1,3	1,15
European ultramafic complexes : (Finero, Hochgrössen, Kraubath)	56	0,1-25	2,0	4,1	0,8	0,65
Barberton Mountain Land : (type locality of Komati Formation)	12	0,4-2,1	1,0	0,5	0,9	0,9
Barberton Mountain Land+ : (type locality of Komati Formation)	9	0,9-1,6	1,2	0,7	1,2	1,1
Pietersburg Belt (Eersteling area)	35	0,3-20,2	2,8	4,4	1,5	1,45
Belingwe Belt (Ngesi River area)	51	0,1-372	18,6	53,8	3,1	1,95

+ Data obtained from Anhaeusser *et al.* (1975)

basaltic rocks from 3,2 ppb Au (Jones, 1969) to 17,4 ppb Au (Boyle, 1979). A comparison of the above values with the data obtained during this study indicates that regional differences in gold abundance grossly affect the tabulated mean values. This is due to the fact that the data used for the tabulations were not systematically collected on the earth's surface but are strongly biased by regional sampling. For instance, the geometric mean value (G) of our data from southern Africa is 2,1 ppb Au ($\bar{x}$ = 10,8 ppb Au), whereas the geometric mean value for samples restricted to South African greenstone belts lies at 1,3 ppb Au ($\bar{x}$ = 2,4 ppb Au), and for the Barberton Mountain Land an even lower geometric mean value of 0,9 ppb Au ($\bar{x}$ = 1,0 ppb Au) was found. Estimations of crustal abundance values of gold are thus probably of little avail and it is suggested that gold abundance values should only be given for restricted petrogenetically well-defined areas.

B. Mineralogical Siting of Gold

Gold is regarded as a distinctly siderophile element. It also has a weak chalcophile nature as it combines with Te, Bi, and Sb but only forms one naturally occurring sulphide, the double sulphide uytenbogaardite, Ag_3AuS_2 (Barton *et al.*, 1978). The lithophile character of gold is extremely weak. Consequently, gold concentrations in silicates are low and one can assume that in silicate minerals the precious metal preferentially occurs in defect structures of the crystal lattices. Boyle (1979) gives gold contents in silicates ranging from 0,2 to 924 ppb, and he suggests that in dark silicates, which often have higher gold contents, gold may be present to a large extent in microscopic inclusions of pyrite and/or magnetite.

All available data on gold contents of minerals indicate that sulphide minerals possess pronounced higher gold contents than silicate minerals (Gottfried *et al.*, 1972; Tilling *et al.*, 1973; Vincent and Crocket, 1960; Crocket, 1974; Boyle, 1979; and others). This demonstrates that the chalcophile character of gold is much greater than its lithophile nature.

In sulphide minerals (including those which occur as accessories in the igneous rocks investigated — pyrite, arsenopyrite, pyrrhotite, and chalcopyrite), the siting of gold is disputed (Ramdohr, 1975). The nature of gold in pyrite has been studied in some detail and Kurauti (1941) described solid solution of gold in pyrite and reported that up to 2000 ppm Au can be substituted in the pyrite lattice. Ramdohr (1975) and others, have, however, suggested that much of the gold in pyrite is present in a submicroscopic particulate form. McPheat *et al.* (1969) concluded that gold in natural and auriferous pyrite is in solid solution up to a concentration of 1 ppm, and that the balance of the metal is finely dispersed as grains throughout the mineral. With the aid of an electron microscope, Iwanoff (1951) showed that in pyrite from Fiji, containing up to 1000 ppm Au, the precious metal formed submicroscopic inclusions with diameters up to 0,01 microns.

For arsenopyrite, Clark (1960) suggested solid solution of gold of approximately 1000 ppm, and Boyle (1979) reported a gold solubility in chalcopyrite of about 1,5 per cent at 600°C. Much of this gold is exsolved on cooling. Very little is known about the solubility of gold in pyrrhotite but Barton (1969) assumed that it is perhaps less than 1000 ppm Au.

The gold content of magnetite may attain levels as high as 50 ppb (Fominykh and Znamenskii, 1974). Gold apparently substitutes for iron in the magnetite since free gold has rarely, if ever, been observed (Boyle, 1979). For chromite, gold contents of less than 10 ppb have been reported in the literature.

The data from southern Africa revealed no correlation between gold and the modified Larson factor as well as between Au and Mg, Cr, Co, and $Ni_{tot.}$ (the latter being typical major and trace elements in the rock-forming minerals of mafic and ultramafic rocks, Table 7). This conspicuous lack of correlation is also borne out by the factor analysis which indicated an absence of fundamental relationships between Au and the other elements investigated (Table 8), and consequently rules out a close systematic relationship between Au and the rock-forming minerals.

The basaltic and peridotitic rocks investigated contain very rare, randomly distributed, sulphides and do not reveal apparent differences in their sulphide contents. On the premise that maximum solubility of gold of several hundred parts per million in sulphides was achieved, only 0,1 weight per cent of sulphides would account for a gold content of more than 300 ppb in the bulk samples. This value lies close to the highest gold concentration (372 ppb) observed in this study. If the gold is present in the sulphides as microscopic and/or submicroscopic inclusions, an even lower solubility of gold is required to explain the "excess gold populations" which were found in all gold distributions, irrespective of the mode of data grouping.

Thus, it is suggested that the gold content of the volcanic rocks is almost exclusively determined by the presence of accessory sulphide minerals. As a more hypothetical consideration, some native gold may also be present as submicroscopic particles, although such particulate gold has not been observed in the material investigated. The absence of particulate gold can probably be explained by the following estimation :- if in a one gram sample, assaying 300 ppb Au (a content only once exceeded in this study), *all* the gold is concentrated in a single cube, the cube would measure only 42,6 microns in diameter. This makes it understandable why no intergranular gold particles were observed in the material investigated.

The bimodal gold distribution, found in the southern African data, can best be explained as follows :

(i) gold occurring in solid solution in sulphides and gold hosted in rock-forming minerals represents the "background population" of gold values (81 per cent of the samples), and

(ii) gold occurring as inclusions, as well as in solid solution in sulphides, together with intergranular particulate gold determines the "excess population" of gold contents (19 per cent of the samples).

A close association between gold and sulphide minerals is obvious for the hornblende-peridotite samples from the Finero ultramafic complex (Tables 9a, 9b). Careful microscopic inspection of a large number of phlogopite-peridotite samples from Finero led Oberthür (1978) to conclude that no sulphides are present in this rock type. Eight phlogopite-peridotite samples possess an arithmetic mean ($\bar{x}$) S content of 49 ppm (G = 45 ppm S) (Table 4). The gold data of the phlogopite-peridotite display a normal distribution ($\bar{x}$ = 0,3 ppb Au; s = 0,1 ppb Au; G = 0,3 ppb Au). On probability paper a cumulative normal population defines a straight line and statistical partitioning therefore does not allow extracting an "excess population" (Sinclair, 1978). From this observation it is deduced, that the gold content of the phlogopite-peridotite, lying between 0,1 and 0,5 ppb, is exclusively determined by gold hosted in the lattice of rock-forming silicate and oxide minerals and not in irregularly distributed accessory phases.

Gold contents less than 0,5 ppb, displaying a single normal distribution in a particular rock type, can thus be used as an indication that in such cases no gold is present in accessory sulphide minerals and/or in a particulate form.

It is significant that the partly serpentinized, distinctly sulphide-bearing, hornblende-peridotites from Finero have sulphur contents ranging from 0,008 to 1,05 per cent ($\bar{x}$ = 0,32% S; G = 0,16% S) and gold

contents lying between 0,2 and 3,0 ppb ($\bar{x}$ = 1,1 ppb Au; G = 0,8 ppb Au) (see also Table 4) and do not exhibit a normal distribution.

C. Gold in Ferruginous Chemical Sediments

The comparison of the mean gold contents of ferruginous chemical sediments (N = 34; $\bar{x}$ = 129,9 ppb Au; G = 35,7 ppb Au) with that of volcanic rocks from South Africa (N = 47; $\bar{x}$ = 2,4 ppb Au; G = 1,3 ppb Au) reveals a marked enrichment of gold in the sediments.

Most of the chemical sediments originate from greenstone terranes and belong to the Algoma-type banded iron-formations (Gross, 1966). One sample derives from the Pongola sedimentary basin and six samples from the Witwatersrand basin. These seven samples can be assigned to the Superior-type banded iron-formations (Gross, 1966). Gold mineralization occurring near some of the ferruginous chemical sediments investigated belong to stratigraphically different units and are not genetically related to the sampled horizons.

In greenstone terranes, banded iron-formations are of minor volumetric importance. They are associated with volcanic rocks and with arenaceous and argillaceous sediments. According to Beukes (1973), the silica in these chemical sediments is of volcanic origin and the iron was probably derived through weathering of a source area. A volcanic exhalation source for iron was not, however, ruled out.

For the Pongola and Witwatersrand banded iron-formations, Beukes (1974) hypothesized that both the iron and silica were derived from the weathering of pre-existing rocks, but he also suggested the possibility that the silica in these rocks may have been of felsic volcanic origin.

A grouping of the samples, in accordance with their genetic characteristics, demonstrates that among the ferruginous chemical sediments, those belonging to the Algoma-type possess markedly higher mean gold parameters (N - 23; $\bar{x}$ 0 172,3 ppb Au; G = 84,6 ppb Au) than those belonging to the Superior-type (N = 9; $\bar{x}$ = 21,3 ppb Au; G = 3,9 ppb Au).

These findings agree with those of Boyle (1979) who observed (in samples from the Canadian Shield) that the Algoma-type iron-formations generally contain more gold than the Superior-type iron-formations. The mineralogical siting of gold in chemical sediments is not yet understood. Boyle (1979) noted that the sulphide-facies banded iron-formations are the most enriched in gold. The samples investigated in this study are oxide-facies members containing randomly distributed minor amounts of sulphide minerals, which, by analogy with the volcanic rocks, may be important gold carriers. In spite of the fact that samples from this study have a different composition from the distinctly sulphide-bearing chemical sediments from Kambalda, Australia, investigated by Bavinton and Keays (1978), both sample populations are similar and show comparable gold values.

Bavinton and Keays (1978) proposed three models to explain the gold enrichment in chemical sediments :

(i) a magmatic-exhalative model,

(ii) a sea-floor leaching model, and

(iii) a metamorphic entrapment model.

We feel that combinations of the three models may furnish the best explanation for the presence of elevated gold contents in the banded iron-formations investigated.

D. Metallogenic Implications

As mentioned earlier, many authors are now of the opinion that a large number of Archaean gold lodes formed by metamorphic secretion processes (cf. Introduction). According to this metallogenic concept the gold and its accompanying elements were mobilized by hydrothermal solution from supracrustal source rocks occurring near to the gold veins.

Keays and Scott (1976) concluded from their geochemical investigations of Mid Atlantic Ridge pillow basalts that the suitability of a volcanic rock as a source for epigenetic lode mineralization is largely determined by the presence of gold in any form leachable by migrating solutions, an axiom upon which any secretion theory as a metallogenetic concept hinges.

Generally, sulphides and particulate gold are more easily attacked and dissolved by migrating hydrothermal solutions than rock-forming silicate and oxide minerals. The foregoing discussion on the mineralogical siting of gold indicated that in volcanic rocks and chemical sediments sulphide minerals and/or particulate gold are the most important gold carriers. Supracrustal rocks thus contain gold readily available to metamorphic secretions. A rough estimation, based on gold solubility data in sulphides, shows that gold concentrations of up to 5 ppb in a rock sample can be accounted for by sulphide contents of less than 0,002 weight per cent. Such low contents of sulphide minerals are usually present in supracrustal rocks suggesting that, in a general sense, *all* supracrustal rocks must be considered as suitable source rocks for gold veins formed by metamorphic secretion processes. In a volume of 1 km^3 of supracrustal rocks, containing one part per million gold associated with sulphides, about 3 t of gold would be available for ore formation! It should be noted here, that in this study only one rock type, namely the Finero phlogopite-peridotite (significantly one of the few intrusive rocks investigated) was found to be free of sulphides, and therefore could be assigned to the class of non-suitable source beds for gold.

Because practically all supracrustal rocks are suitable gold sources in greenstone terranes or in other areas predominantly underlain by supracrustal rocks, the "source bed" carries little weight for the control of gold mineralization, and it follows that the distribution and the occurrence of gold deposits, formed by secretion processes, depend on other more powerful metallogenetic factors. This means, that the conspicuous abundance of gold lodes in Precambrian greenstone belts cannot be based on the hypothesis that early Precambrian supracrustal rocks are better gold sources than younger equivalent rocks. It also means that the level of the gold content in a source rock cannot be regarded as the determining factor for the occurrence of metamorphically

formed gold veins. This is borne out by the present study which has shown that the rocks with the highest gold level, i.e. those from the Upper Greenstones of the Belingwe Belt, are not associated with gold mineralization. This observation is in agreement with the findings of Tilling *et al.* (1973) who pointed out that the concentration of gold in possible source rocks is a trivial factor for the formation of gold vein deposits. Furthermore, it follows that a high availability of gold in a source rock need not guarantee the formation of metamorphically formed gold veins, a feature that restricts the use of "background gold levels" in rocks as an indicator in geochemical gold exploration. However, it seems obvious, that if vein deposits were emplaced by secretion processes, the availability of leachable gold, would, to a certain extent, determine the ore grade of such deposits.

Thus the questions remain as to what factors control the abundance of epigenetic gold deposits in the Precambrian, and why are certain Precambrian areas devoid of gold veins whereas others are enriched in gold veins? Of the Precambrian samples investigated only those from the Eersteling area of the Pietersburg Belt originated from a terrane which had suffered substantial dynamothermal metamorphism. This area contains many economic gold quartz-veins (Saager and Muff, in press). It is therefore suggested that the metamorphic overprint is the most important factor controlling the formation of gold veins by secretion processes. Of almost equal importance is the presence of suitable traps — generally faults and shear zones forming dilatant structures in competent rock types — into which the gold, sulphides and associated gangue components can be drawn and concentrated. Such traps must obviously have an intimate spatial relationship to the potential gold source.

Anhaeusser (1973) considered that the development of Archaean greenstone terranes differed significantly from the processes leading to the formation of younger orogenic belts. According to the model he proposed Archaean granite-greenstone complexes evolved on a primitive unstable crust through gravity induced folding and faulting of supracrustal rocks and, as a result of widespread intrusion of diapiric plutons. These plutons caused, or were intimately related to, regional metamorphism and generated hydrothermal secretions. At the same time, higher order folds, faults, fractures and shear zones were developed in adjacent competent rock types forming sinks for the components of the mineralizing solutions (Viljoen *et al.*, 1969; Boyle, 1961, 1979; Anhaeusser, 1976; and others).

The occurrence of source rocks moderately metamorphosed during syntectonic granite intrusions and the presence of suitable dilatant zones in well-preserved, spatially restricted areas *not* affected by large-scale low angle thrusts, intense metamorphism, and other processes akin to typical Alpine-type mountain building, are unique to the Archaean greenstone belts. This unique preservation of source rocks and nearby structures (the latter suitable for being mineralized in competent rock types) is regarded as the key factor for the pronounced abundance of epigenetic gold veins in Precambrian shield areas. Other factors that are often considered, such as elevated gold contents in potential source beds, the composition of the source rocks, or the chemistry of the leaching solutions, are assumed to be of lesser importance for the control of metamorphically formed gold mineralization.

Considering the ferruginous chemical sediments, it can be assumed that there are no apparent reasons why the gold in these beds is not accessible to migrating leaching solutions during periods of regional metamorphism. Therefore, these rocks, as well as their enclosing volcanics, form favourable sources for gold. In the Marabastad Goldfield of the Pietersburg greenstone belt, Saager and Muff (in press) report gold occurrences associated with thin quartz stringers and veinlets in strongly deformed oxide facies banded iron-formations. The authors propose a direct genetic link between the gold veins and the enclosing ferruginous chemical sediments; they also emphasize that supergene enrichment of gold may have played an important role in the formation of these vein-type gold ores.

VI. SUMMARY AND CONCLUSIONS

1. Ninety-eight samples of mafic and ultramafic volcanics from early Precambrian greenstone terranes of southern Africa were analysed for Mg, Al, Cr, Fe, Co, $Ni_{tot.}$, and Au. The gold values range from 0,1 to 372 ppb and exhibit a positively skewed, quasi log-normal distribution.

2. As a comparative population, 56 samples from Palaeozoic European ultramafic complexes were investigated for the same suite of elements. The gold data also follow a quasi log-normal distribution, but show a smaller range from 0,1 to 25 ppb Au.

3. The samples from southern Africa exhibit no correlation between gold and major as well as trace elements indicating a lack of relationship between gold and rock-forming minerals. Factor analysis of the variables Au, Mg, Cr, Fe, Co, and $Ni_{tot.}$ also revealed the independent behaviour of gold.

4. Graphical partitioning of bimodal probability curves, which were plotted for various data sets of gold contents (the latter grouped according to geographical and petrographical aspects) demonstrates the presence of "background populations" and "excess populations". The latter are made up of a minor proportion of conspicuously high gold values lying above a statistically determined threshold value.

5. Ferruginous chemical sediments of the oxide facies from South Africa possess gold concentrations lying between 0,5 and 667 ppb. The mean gold content of the chemical sediments is decidedly higher than the mean gold content of associated volcanic rocks. Ferruginous chemical sediments belonging to the Algoma-type were found to contain more gold than the younger ones of the Superior-type.

6. In spite of careful microscopic investigations, no gold was detected in any of the samples studied. Thus, the siting of gold had to be indirectly inferred from mineralogical and chemical data. It is suggested that accessory sulphides are the most important gold carriers in volcanic rocks and probably also in the chemical sediments studied. Some native gold probably also occurs in intergranular particulate form. However, the extremely small size of such gold particles makes their detection in polished sections unlikely. Association of gold with sulphide minerals is reflected in the data set by the occurrence of a "background" and an "excess population". A single, normally distributed population of gold contents, ranging from 0,1

to 0,5 ppb, was found in sulphide-free phlogopite-peridotite samples. From this fact it is suggested that rock-forming silicates and oxides, hosting gold in their crystal lattices, contribute usually less than 0,5 ppb of the gold in rocks.

7. Geologically, geochemically, and petrographically similar volcanic rocks, collected from various areas, possess different gold levels, thus indicating the occurrence of regional differences in gold abundance. These differences may reflect inhomogeneities of gold abundance in the upper mantle. On the other hand, the assayed mafic and ultramafic volcanics do not differ in their gold levels.

8. Gold present in sulphides and/or particulate form is more readily accessible to metamorphic hydrothermal solutions than gold locked up in the lattice of rock-forming silicate and oxide minerals. Most supracrustal rocks contain sulphide minerals and thus are a suitable source of gold for the formation of gold veins as required by the metamorphic secretion theory.

9. In areas underlain by supracrustal rocks, metamorphic overprinting of these rocks is considered to be the key factor leading to the eventual formation of gold veins by secretion processes, whereas the gold levels of potential source rocks appear to have little influence on the formation of gold lodes. The conspicuous preponderance of vein gold deposits in Archaean granite-greenstone terranes is attributed to the unique evolution of such terranes. This evolution resulted in the preserved simultaneous occurrence of source rocks for gold, of intrusive granites which caused or enhanced the metamorphic secretion processes and, of dilatant structures suitable for ore deposition in well-defined, spatially restricted areas.

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